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CONTENTS

	Pages
Interpretation of the configuration of the Aleutian Ridge. By Olcott Gates and William Gibson	127–146
Petrology and weathering of the Iron Dike, Boulder and Larimer counties, Colorado. By Ernest E. Wahlstrom	147–164
Upper Croixan stratigraphy, Upper Mississippi Valley. By C. A. Nelson	165–184
Pegmatites in West Greenland. By Hans Ramberg	185–214
Geology of Isabela (Albemarle) Island, Archipielago de Colon (Galapagos). By A. F. Banfield, Chas. H. Behre, Jr., and David St. Clair	215–234
Short Notes	
Late Keweenawan or early Cambrian glaciation in Upper Michigan? By Wm. C. Gussow	235–236
Late Keweenawan or early Cambrian ice shove in upper Michigan—a reply. By Raymond C. Murray	237–238

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Structure of Lopoliths. By H. D. B. Wilson

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INTERPRETATION OF THE CONFIGURATION OF THE ALEUTIAN RIDGE

BY OLCOTT GATES AND WILLIAM GIBSON

ABSTRACT

Data from a contour map of the submarine topography surrounding the Near Islands, from the companion map of the Rat Islands by Gibson and Nichols (1953), and from the geology of the western Aleutian Islands suggest that the submarine topography reflects the structure of the western part of the Aleutian Ridge.

Four principal topographic provinces are recognized: (1) The Crest of the Aleutian Ridge contains the Aleutian Islands, the Insular Shelf at depth ranging from present shore lines to 70 fathoms, and the Ridge Shelf at a depth of 100 to 500 fathoms, all apparently the result of subaerial and marine erosion since the middle Tertiary and of glaciation in the late Pleistocene. (2) The Insular Slopes form the sides of the Aleutian Ridge. The North Insular Slope is a long, steep, linear scarp that probably marks a major fracture in the earth's crust. The South Insular Slope appears to be a broad, faulted and warped arch containing numerous steep-sided linear sea valleys and canyons. Many of these traverse the south slope at an angle to the maximum regional gradient, and several line up with observed faults on the islands. These linear topographic features probably mark fault zones. (3) The Aleutian Bench is a prominent step in the general slope from the islands to the Aleutian Trench, and its inside edge may be the trace of a thrust fault. (4) The arcuate Aleutian Trench has a steep north side, a flat floor at a depth of about 4000 fathoms, and a south side containing an en echelon topographic pattern. The Trench perhaps marks a major thrust zone dipping north beneath the Aleutian Ridge.

A structural interpretation of the submarine topography suggests that the western part of the Aleutian Ridge is an arched and faulted asymmetrical wedge bounded by a northward-dipping normal fault on the north and by a northward-dipping zone of reverse faults on the south. Formation of this wedge probably began with major uplift and faulting of the western Aleutian area during the middle Tertiary, and the many earthquakes and active volcanoes in the Aleutian arc today indicate that deformation is still continuing.

CONTENTS

	Page
Introduction	128
Geology of the Western Aleutian Islands	128
Submarine topography	130
General statement	130
Crest of the Aleutian Ridge	132
The islands	132
Insular Shelf	132
Ridge Shelf	133
Origin of the Crest of the Aleutian Ridge	134
Insular Slopes	138
North Insular Slope	138
South Insular Slope	138
Origin of the Insular Slopes	139
Aleutian Bench	143
Aleutian Trench	143
Role of submarine erosion and deposition	144
Structure of the western part of the Aleutian Ridge	144
References cited	145

ILLUSTRATIONS

Figure	Page
1. Index map	128
2. Fracture pattern in the Near Islands	130
3. Locations of submarine topographic units	131
4. Locations of profiles and fathograms of Near Islands Platform	132
5. Profiles of Near Islands Platform	133
6. Profile southeast from Little Sitkin Island	134
7. Fathogram of Near Islands Platform	135
8. Profiles of Aleutian Ridge	136
9. Profiles of terraces now being cut in bedrock	137
10. Cross-sectional and longitudinal profiles of five sea valleys	139
11. Fathograms from head of Abraham Sea Valley	140
12. Distribution of supposed faults	141
13. Diagrammatic cross section of western part of the Aleutian Ridge and Aleutian Trench	145

Plate	Facing page
1. Configuration of Aleutian Ridge, Near Islands group	127
2. Views on Attu and Agattu islands	136

TABLE

Table	Page
1. Summary of the geology of the Near Islands	129

Introduction

This paper presents a contour map of the submarine topography in the vicinity of the Near Islands, the westernmost islands of the Aleutian Island arc. The map joins the map of the Rat Islands by Gibson and Nichols (1953, Pl. 1) and extends the contouring westward to 170° East Longitude.

The writers discuss the origin of the submarine topography on the basis of data from the Near Islands map, the Rat Islands map, and the geology of the western Aleutians. They amplify, and in part modify, the conclusions of Gibson and Nichols (1953) to whom post-war geologic studies were not available.

Detailed surveys from 1946 to 1951 by the U. S. Coast and Geodetic Survey provided the basic hydrographic data for the contour map. The geologic data stem from a reconnaissance of the western Aleutians from 1946 to 1953 by the U. S. Geological Survey in co-operation with the U. S. Department of Defense. The geology of the western Aleutians has been described by Capps (1932); Sharp (1946); Coats (1950; 1951; 1952); Wilcox, Gates, and Powers (1952); Todd (1953); Gates, Fraser, and Snyder (1954).

Gibson is primarily responsible for the contour map. Before the two authors pooled their efforts, Gates had contoured independently about two-thirds of the area and in this large area of duplicated work the two maps were almost identical. This identity reduces, but does not necessarily eliminate, the possibility that individual interpretation has significantly distorted the under-water topography.

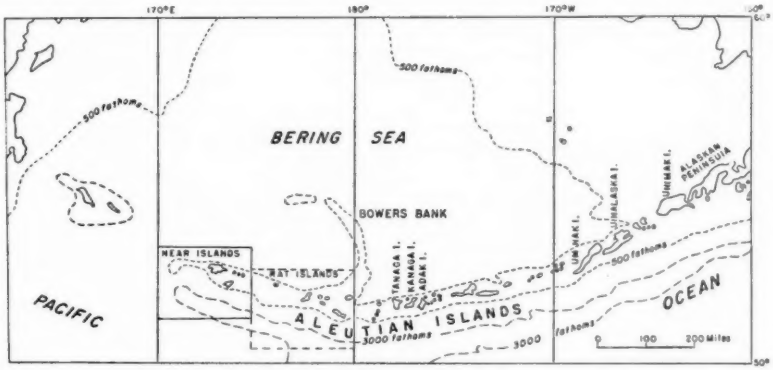

FIGURE 1.—INDEX MAP

Area of Plate 1 is outlined by solid line, area of Rat Islands outlined by dashed line (Gibson and Nichols, 1953, Pl. 1).

Gibson and Nichols (1953) describe in detail the submarine topography around the Rat Islands and also discuss hydrographic survey methods and accuracy. In this paper frequent reference will be made to the map of the Rat Islands (Gibson and Nichols, 1953, Pl. 1).

The authors gratefully acknowledge the constructive criticism given this paper by H. A. Powers, R. E. Wilcox, George Fraser, and George Snyder, of the U. S. Geological Survey, Philip C. Scruton, H. W. Menard, K. O. Emery, and also the encouragement and assistance given by Rear Admiral R. F. A. Studds, Director, U. S. Coast and Geodetic Survey, in providing scale reproductions of the many unpublished survey sheets.

Geology of the Western Aleutian Islands

Table 1 summarizes the geology of the Near Islands and Rat Islands.

The unconformity separating lower to middle Tertiary rocks from upper Tertiary rocks has been mapped on Kiska, Adak, Umnak, and Unalaska islands, almost the length of the

TABLE 1.—SUMMARY OF THE GEOLOGY OF THE NEAR ISLANDS

Age	Sedimentation		Igneous activity		Tectonic activity
	Near Islands	*Rat Islands*	*Near Islands*	*Rat Islands*	
Late Tertiary and Quaternary	Erosion, glaciation, local subaerial sediments		None	Continued activity of stratovolcanoes Buldir, Kiska, Segula, Little Sitkin, and Semisopochnoi	Normal faulting at a reduced scale continuing to present; many earthquakes, local warping, several fluctuations in sea-level relative to land
	Attu Coarse subaerial conglomerate and sandstone	Shallow-water, nearshore conglomerate and sandstone	*Attu* Subaerial hornblende andesite lava flows	Shallow-water lava flows and pyroclastic rocks associated with Kiska stratovolcano	
					Unconformity: Major uplift and erosion, much normal faulting, some folding
Early to middle Tertiary	Marine conglomerate overlain by banded argillite, sandstone, and limestone, all deposited in shallow water	Siltstone, sandstone, conglomerate. Eocene to Miocene pelecypods, Eocene and Oligocene Foraminifera	Intrusion of gabbro, quartz diabase, albite granite, diorite in sills, dikes, small plutons; calc-alkalic circum-Pacific type rocks		
	A few fossils				Unconformity
Mesozoic(?)	Bedded and laminated chert, argillite, graywacke, sandstone, and conglomerate; marine, unfossiliferous		Basaltic and spilitic pillow lavas, basaltic and keratophyric pyroclastic beds including breccia and coarse agglomerate		Obscure—rapid facies changes, much lensing, scouring, contemporaneous deformation of beds suggest many local centers of deposition and considerable relief

Aleutian arc, and marks a turning point in the geologic history. Mesozoic (?) and lower to middle Tertiary rocks beneath the unconformity are for the most part a spilitic suite of marine sediments, marine pyroclastics, and pillow lavas. Upper Tertiary and Quaternary rocks above the unconformity are primarily subaerial or very shallow-water sediments, pyroclastics, and lava flows associated with the many calc-alkalic stratovolcanoes strung out along the northern edge of the Aleutian arc.

Scarcity of fossils prevents accurate dating of the erosion, but middle Tertiary is probable. In the Near and Rat islands, the youngest rocks cut by the unconformity carry Eocene and Oligocene Foraminifera (Todd, 1953) and a few early Tertiary pelecypods that may be as young as Miocene. The stratovolcanic rocks overlying the unconformity appear to be upper Tertiary to Recent. Shallow-water pyroclastic beds associated with the beginning of Kanaga and Tanaga stratovolcanoes yielded Pliocene and Pleistocene fossils. Glaciation of several of the stratovolcanoes suggests that they were approximately their present size and shape during late Pleistocene and thus had been presumably growing for considerable time prior to glaciation.

The cutting of the middle Tertiary erosion surface followed a major orogeny accompanied by extensive folding and faulting as well as intrusion of igneous rocks. Deformation on a reduced scale faulted upper Tertiary rocks, and the Aleutian arc is still tectonically restless. Earthquakes are numerous, volcanoes active, and fault scarps on many islands offset glacial and postglacial surficial deposits and lava flows of Recent age.

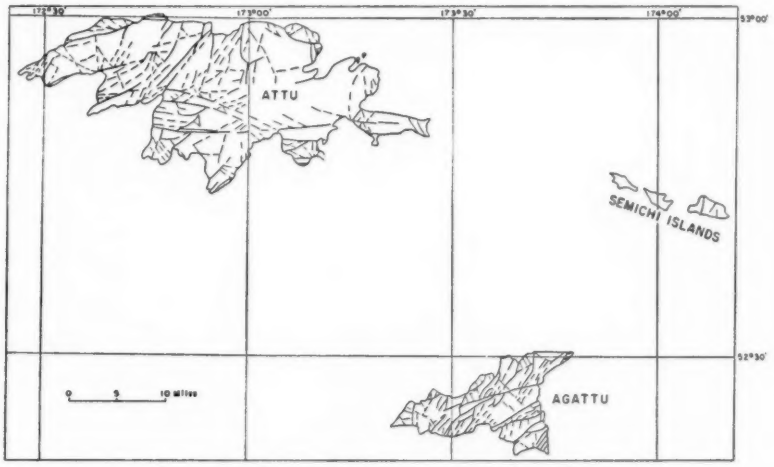

FIGURE 2.—FRACTURE PATTERN IN THE NEAR ISLANDS
Solid lines are faults found during mapping. Dashed lines are fractures inferred from aerial photographs

Normal faults and open folds characterize the structure of western Aleutian islands. Figure 2 shows the complex pattern of fracturing in the Near Islands. Some of the faults have displacements of several thousand feet, and movement was vertical along some faults and transverse along others. The regional dip of the rocks is steep to the north along the north coast of Attu and gentle to the south on Agattu.

SUBMARINE TOPOGRAPHY

General Statement

The Aleutian Islands form the highest parts of the arcuate Aleutian Ridge separating the Bering Sea from the Pacific Ocean. This ridge extends more than 1000 miles westward from the Alaskan Peninsula and rises 12,000 feet above the floor of the Bering Sea and 18,000

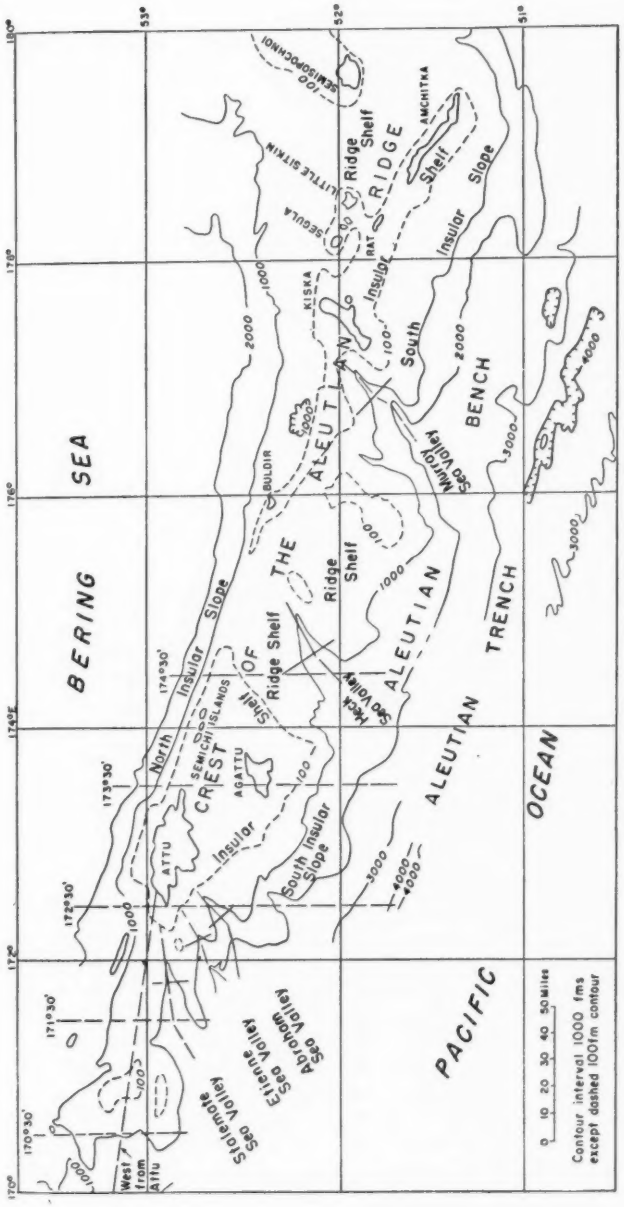

FIGURE 3.—LOCATIONS OF SUBMARINE TOPOGRAPHIC UNITS

Long-dashed lines show locations of cross sections of Aleutian Ridge (Fig. 8) and of cross sections and profiles of sea valleys (Fig. 10)

feet above the floor of the Pacific Ocean. The southern convex side is bordered by a long narrow trough, the Aleutian Trench, which descends to a depth ranging from 21,000 to 25,000 feet below sea level.

Attu produced a rugged alpine terrain (Pl. 2, fig. 1).

Insular Shelf.—The Insular Shelf is a prominent platform that indents the stratovolcanoes, extends from south of Amchitka to northwest

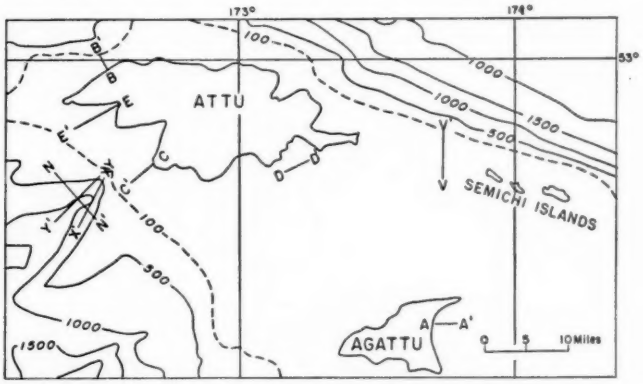

FIGURE 4.—LOCATIONS OF PROFILES (A-A′, B-B′, C-C′, D-D′, E-E′, Fig. 5) AND OF FATHOGRAMS (V-V′, Fig. 7; X-X′, Y-Y′, Z-Z′, Fig. 11) OF NEAR ISLANDS PLATFORM

This paper is concerned primarily with the western 450 miles of the Aleutian Ridge and the Aleutian Trench. In this area four main submarine topographic provinces are differentiated (Fig. 3). In order of increasing depth, the main provinces and their subordinate units are: (1) Crest of the Aleutian Ridge including the islands, Insular Shelf, and Ridge Shelf; (2) Insular Slopes including the North Slope and the South Insular Slope; (3) Aleutian Bench; and (4) Aleutian Trench.

Crest of the Aleutian Ridge

The islands.—The islands of Buldir, Kiska (in part), Segula, Little Sitkin, and Semisopochnoi are stratovolcanoes, which in spite of considerable glacial erosion retain their volcanic constructional shape. Attu, Agattu, the Semichi Islands, Rat, Amchitka, and the southern part of Kiska lack young stratovolcanoes and are composed of pre-middle Tertiary rocks and subordinate amounts of upper Tertiary coarse clastic sediments and subaerial lava flows. They probably owe their height to faulting and warping and since middle Tertiary have been extensively eroded. Vigorous glaciation modified the preglacial topography and on

of Buldir, and includes Tahoma Reef as an outlier (Gibson and Nichols, 1953, Pl. 1). Around the Near Islands, it forms a triangular platform with Ingenstrom rock at the east corner, Agattu Island at the south corner, and Attu Island at the west corner (Pl. 1). Stalemate Bank, west of Attu, is also part of the Insular Shelf.

The 70-fathom contour line approximately outlines the edge of the Insular Shelf around the Near Islands. In the Rat Islands the depth of the edge is also about 70 fathoms (Gibson and Nichols, 1953, p. 1182) but is not shown because of absence of 10-fathom contour intervals. The upper limit of the shelf is considered to be present shore line around the islands.

The following discussion concerns primarily the Insular Shelf around the Near Islands because more data are available than from elsewhere in the western Aleutians. The topography is delineated by 10-fathom contours (Pl. 1), and a survey of the bottom sediments has been made (Scruton, 1953). The Insular Shelf around the Near Islands is designated the Near Islands Platform to distinguish it from the Insular Shelf elsewhere in the Aleutians.

The topography of the Platform has three principal units:

(1) The slope around the islands from about 40 fathoms to the present shore lines is a submarine extension of the general subaerial shape of the islands, and around Attu includes extensions of the main glaciated valleys. In many places, it consists of a series of steps (Fig. 5, A-A′, B-B′, C-C′). However, in the extensions of the broad valleys of Attu, such as Etienne, Abraham, and Holtz, the profile is smooth (Fig. 5, E-E′). Steplike profiles are also around the volcanic islands (Fig. 6).

(2) The zone of irregular hummocky topography southeast of Massacre Bay on Attu begins at a depth of about 40 fathoms and includes numerous reefs and shoals (Fig. 5, D-D′; Fig. 7).

(3) A broad, relatively flat outer shelf extends from the 70-fathom contour to about the 40-fathom contour. Its low relief is interrupted by several large depressions having a maximum depth of about 90 fathoms and a few low rounded mounds and knobs (Fig. 7; Pl. 1).

Scruton (1953) has made a detailed survey of the unconsolidated sediments on the Near Islands Platform. Bedrock, or boulders too large to be sampled by Scruton's equipment, and scattered patches of sand underlie most of the flat outer shelf and the steplike rise around the islands. Mixed pebbles, granules, and fine sediment fringe the southwest edge of the outer shelf and extend down the South Insular Slope. The depressions contain a mixture of sand and planktonic debris. Cobbles and gravel are abundant in the area of hummocky topography, and narrow tongues of sand extend from the mouths of several of the main streams on Attu.

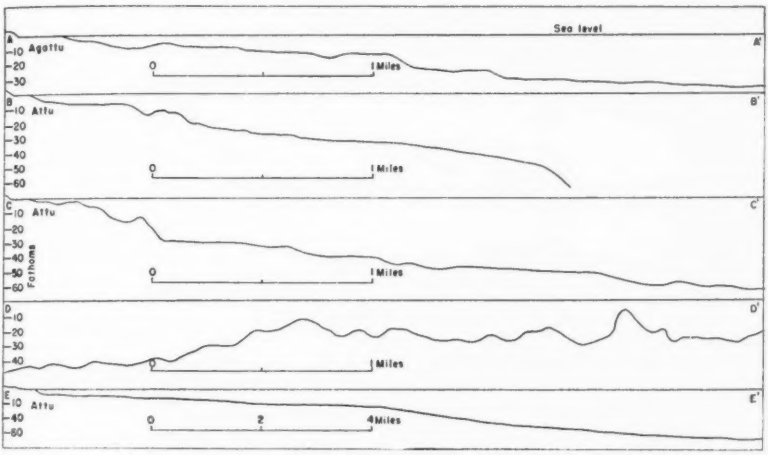

FIGURE 5.—PROFILES OF NEAR ISLANDS PLATFORM

A-A′, B-B′, C-C′ are of the steplike rise around the islands. D-D′ is of hummocky topography southeast of Massacre Bay. E-E′ is of submerged extension of Etienne Valley on Attu. Profiles taken from continuous sounding runs plotted on original boat sheets. *See* Figure 4 for locations.

Ridge Shelf.—The Ridge Shelf forms the submerged top of the Aleutian Ridge in several places where the Insular Shelf is absent. It lies at a depth ranging from about 100 to 500 fathoms. It is present west of Attu (Pl. 1), between Agattu and Buldir (Pl. 1), and northeast of Amchitka (Gibson and Nichols, 1953, Pl. 1).

The Ridge Shelf is less persistent in aerial extent, has considerably more relief, and is less consistent in general depth than the Insular Shelf. Between Agattu and Buldir, the Ridge Shelf is between 100- and 500-fathom contours and in part shows considerable linear topography, a subdued extension of the linear topography on the northwest side of Heck Sea Valley. The inner part has little relief

and abuts against a submerged cliff. South of Buldir, the Ridge Shelf is about 400 fathoms deep and is bordered on the east by an irregular submarine cliff northwest of Tahoma Reef

began erupting at the same time, an assumption supported only by their position above the unconformity and by meager paleontological data.

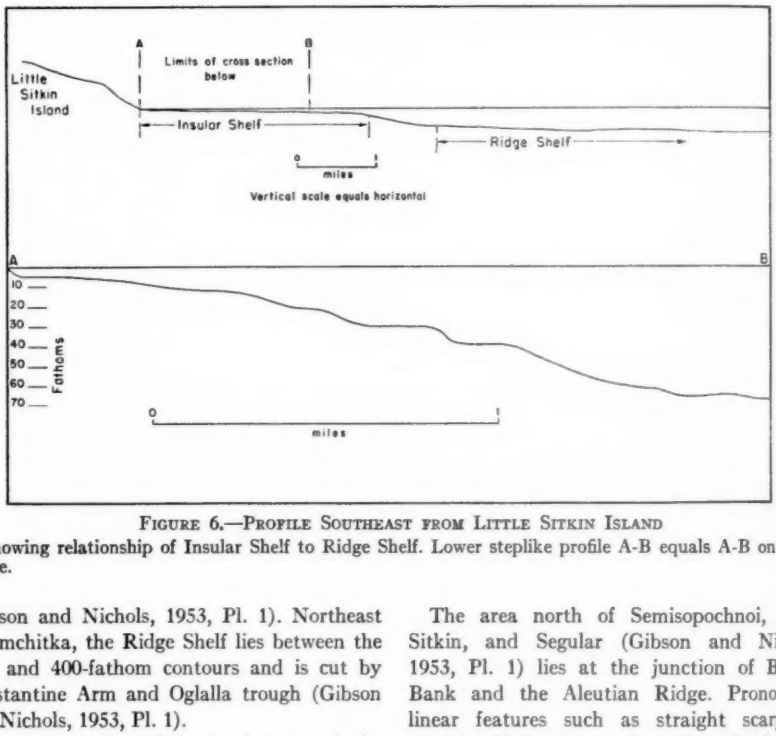

FIGURE 6.—PROFILE SOUTHEAST FROM LITTLE SITKIN ISLAND

Showing relationship of Insular Shelf to Ridge Shelf. Lower steplike profile A-B equals A-B on upper profile.

(Gibson and Nichols, 1953, Pl. 1). Northeast of Amchitka, the Ridge Shelf lies between the 300- and 400-fathom contours and is cut by Constantine Arm and Oglalla trough (Gibson and Nichols, 1953, Pl. 1).

The comparative flatness of parts of the Ridge Shelf with attendant submerged cliffs, subdued linear topography in some places, and the various depths of the Shelf suggest remnants of a wave-cut platform that has been warped and faulted. This supposed wave-cut platform may be the submarine counterpart of the middle Tertiary unconformity. The topography in the vicinity of Segula, Little Sitkin, and Semisopochnoi (Fig. 6; Gibson and Nichols, 1953, Pl. 1) suggests that these stratovolcanoes are built on the Ridge Shelf. Kiska and other volcanoes farther east overlie the middle Tertiary unconformity. Thus this surface and the Ridge Shelf may once have been respectively the subaerial and submarine portions of one surface of erosion. This conclusion assumes that most of the large stratovolcanoes

The area north of Semisopochnoi, Little Sitkin, and Segular (Gibson and Nichols, 1953, Pl. 1) lies at the junction of Bowers Bank and the Aleutian Ridge. Pronounced linear features such as straight scarps or straight-sided canyons that might lend themselves to a structural interpretation are absent. Bowie Canyon has a symmetrical V-shaped cross section, a graded floor with accordant tributaries, a dendritic pattern, and follows the regional slope, all characteristics typical of continental slope canyons for which an origin by some type of erosion seems indicated (Daly, 1942, p. 136; Shepard, 1948, p. 207–249; 1951; 1952; Kuenen, 1950, p. 485–526; 1953, p. 1299–1304; Crowell, 1952).

Origin of the Crest of the Aleutian Ridge.—Geologic and topographic evidence suggests that the Crest of the Aleutian Ridge, except for the stratovolcanoes, is a surface of erosion cut across the top of the Ridge. The cross profiles (Fig. 8; Gibson and Nichols, 1953, p. 1180–1181) show a marked change of slope

from the relatively steep Insular Slopes to the relatively horizontal surface of the Crest. The sea valleys and scarps of the Insular Slopes, which apparently are topographic expressions of fault zones, are abruptly truncated along the edge of the Crest; and similarly, faults on the islands have no topographic expression on the Near Islands Platform, which is underlain by bedrock with only scattered patches of sand and gravel. The peaks of Attu and Agattu consist of old (Mesozoic?) spilitic rocks indicating erosion of thousands of feet of lower Tertiary marine sediments and upper Tertiary coarse clastic rocks and subaerial lava flows, preserved only as scattered remnants overlying the spilitic suite. On some of the islands erosion has cut deep enough to expose coarse-grained intrusive rocks and has erased all topographic expression of fault displacements except for small postglacial scarps.

Subaerial, glacial, and marine erosion have all probably contributed to truncation of the Aleutian Ridge, but the relative importance of each is difficult to assess. The middle Tertiary surface is the oldest evidence of widespread subaerial erosion. Coarse subaerial sediments overlying this surface on Attu and rounded pebbles and cobbles in the tuffs of Amchitka and Rat islands suggest subaerial erosion during late Tertiary time. The unconformity between Mesozoic (?) and early to middle Tertiary sediments on Attu indicate an older period of erotion, but whether this erosion was local or widespread and whether it contributed to the present truncated appearance of the Aleutian Ridge are not known.

Pleistocene glaciation has left a definite imprint on the topography of the Crest. The islands have been vigorously glaciated. Glaciation also helped produce the topography and bottom sediments of the Near Islands Platform. Scruton (1953) concludes that a Pleistocene icecap covered most of the Platform. The ice scoured the depressions, rounded the knobs, and deposited glacial moraine to form the hummocky area of abundant pebbles and cobbles southeast of Massacre Bay. Ice rafting probably produced the mixed sediments along the southwest edge of the Platform.

However, glacial erosion probably removed only a fraction of the total rock eroded during

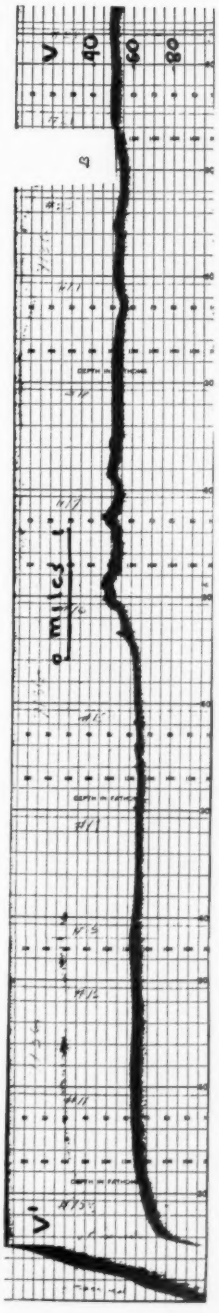

FIGURE 7.—FATHOGRAM OF NEAR ISLANDS PLATFORM

Between Attu Island and Semichi Islands. To left (north) is smooth outer edge of the Platform and beginning of the North Insular Slope. To right is typical hummocky topography believed to consist of glacial deposits. Vertical scale about 15× horizontal. *See* Figure 4 for location

truncation of the Aleutian Ridge. An amount of rock equal to the volume of space between the edge of the Near Islands Platform and the valleys and produced an alpine topography, but have not reduced the islands to low relief. The absence of depressions and troughs ex-

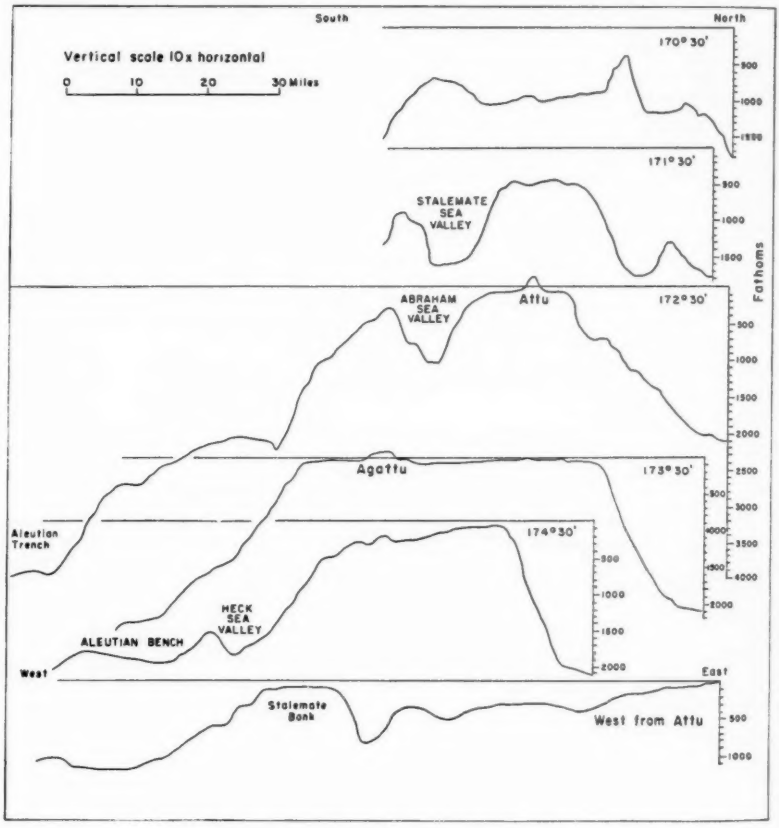

FIGURE 8.—PROFILES OF ALEUTIAN RIDGE
See Figure 3 for locations.

Attu's highest peak is approximately 500 cubic miles, or a layer 1000 feet thick covering the entire Platform. That glacial erosion alone could have removed such a large volume of rock does not seem likely (Flint, 1947, p. 76–88). In the mountainous islands of Attu and Agattu, a terrain in which glacial erosion would be vigorous, glaciers have broadened and deepened tending across the lip of the Platform and down the Insular Slopes suggests that the ice was either almost afloat or relatively thin along the outer edge of the Platform and thus incapable of much erosion. Furthermore, both Attu and Agattu contain remnants of pre-glacial wave-cut terraces between 200 and 600 feet above sea-level. The glaciers apparently originated

PLATE 2.—VIEWS ON ATTU AND AGATTU ISLANDS

FIGURE 1.—MOUNTAINOUS GLACIATED TERRAIN, ATTU ISLAND
FIGURE 2.—MODERN WAVE-CUT TERRACE, SOUTH COAST OF CHIRIKOF POINT, ATTU ISLAND
FIGURE 3.—PREGLACIAL MARINE TERRACE ON SOUTHERN AGATTU

FIGURE 1

FIGURE 2

FIGURE 3

VIEWS ON ATTU AND AGATTU ISLANDS

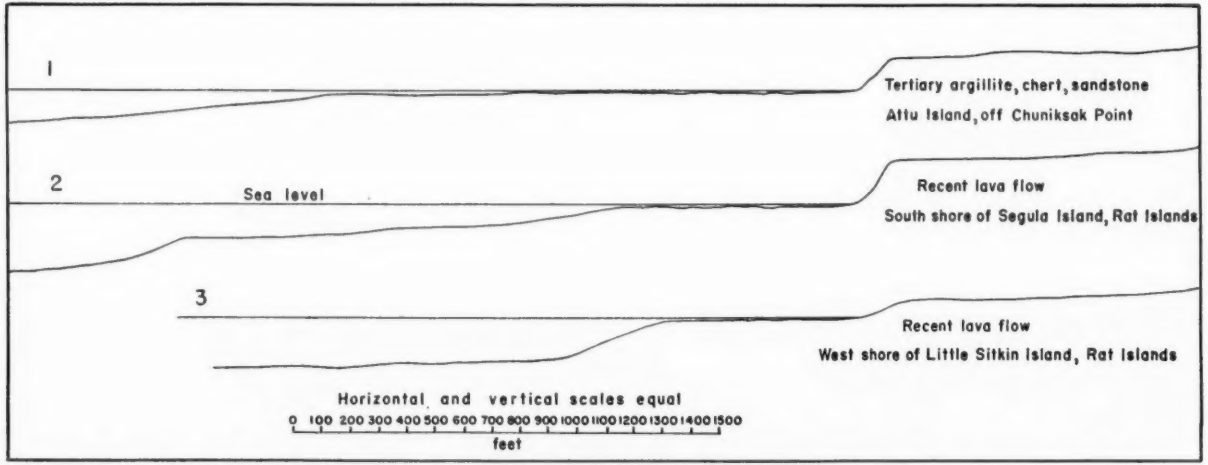
1
Tertiary argillite, chert, sandstone
Attu Island, off Chuniksak Point
2
Sea level
Recent lava flow
South shore of Segula Island, Rat Islands
3
Recent lava flow
West shore of Little Sitkin Island, Rat Islands
Horizontal and vertical scales equal
0 100 200 300 400 500 600 700 800 900 1000 1100 1200 1300 1400 1500
feet

FIGURE 9.—PROFILES OF TERRACES NOW BEING CUT IN BEDROCK

on islands already present and spread across a pre-existing platform.

Marine planation also undoubtedly played an important part in the truncation of the Aleutian Ridge. Some of this planation occurred during the changes in sea level accompanying glaciation. The steplike profiles to a depth of about 40 fathoms around the islands probably indicate wave erosion during the rise of sea level as glaciation waned. Many of the western Aleutian Islands have a discontinuous fringe of shore platforms cut in bedrock at about mid-tide level (Fig. 9; Pl. 2, fig. 2). The profiles of the modern shore platforms resemble those of the submerged steps, and both appear to be parts of the same steplike sequence. This suggests that the slope around the islands from present shore lines to a depth of about 40 fathoms was subjected to wave terracing during postglacial rise of sea level.

During this rise in sea level, some parts of the Platform apparently were protected by remnants of the icecap. Profiles (Figs. 5, 7) of the hummocky topography southeast of Massacre Bay have a morainal aspect unaltered by apparent beaches or terraces. Scruton (1953, p. 42–48) found that this area contains fewer discoid cobbles than do the modern beaches along the shore of Attu. The profile (Fig. 5, E-E′) of the submerged extension of Etienne Valley on Attu shows no steps. During the postglacial rise of sea level, lobes of ice, fed by glaciers moving down the large valley of Attu, apparently remained aground in the hummocky areas and in the submerged extensions of the large valleys. These lobes locally protected the sea bottom from wave erosion while waves were cutting shore platforms on the areas around the islands left unprotected by the general withdrawal of the icecap.

Marine planation appears to be the principal cause of the relatively flat outer part of the Near Islands Platform. No stream drainage pattern is evident from the topography, and submerged extensions of the large valleys on Attu grade to the flat outer shelf of the Platform at a depth of about 40 fathoms. In the Near Islands, extensive wave-cut terraces, now elevated above sea level, testify to several episodes of marine planation and to fluctuations of the strand of considerable magnitude. The southern two-thirds of Agattu is a pre-glacial wave-cut terrace rising from an altitude of about 200 feet on the south coast to 600 feet along the base of the mountain range near the north coast (Pl. 2, fig. 3). Pre-glacial marine terraces on Attu, best exposed on headlands along the south coast, range in altitude from 250 to 500 feet. The flat outer shelf of the Platform probably represents a rather long stand of the sea at a level lower than today. Dietz and Menard (1951, p. 2011) conclude that planation by wave action is restricted to a depth of 5 fathoms or less.

The Near Islands Platform and the rest of the Insular Shelf are not older than late Tertiary nor younger than the latest major (Wisconsin?) glacial maximum in the area; the Insular Shelf indents the stratovolcanoes, and the Near Islands Platform has been glaciated. However, only the youngest major glaciation has left a geologic record. The marine planation responsible for the flat outer shelf may have preceded all Pleistocene glaciation or may have occurred during a lowering of sea level accompanying one or several Pleistocene glacial episodes older than the recorded one.

Insular Slopes

North Insular Slope.—The North Insular Slope extends 90 miles from northwest of Attu to east of Buldir, and is free from major topographic irregularities except for a small assymetrical valley off Holtz Bay. Northwest of Attu, a series of ridges and troughs replaces the single scarp.

The North Insular Slope has a fairly uniform gradient of about 1600 feet per nautical mile. The bottom lies along the 1900-fathom contour, beyond which a narrow apron extends to the flat floor of the Bering Sea at a depth of about 2100 fathoms.

South Insular Slope.—The South Insular Slope is very irregular in topography, has a variety of gradients, and no persistent linear trend. It roughly parallels the curve of the Aleutian Ridge as far west as Agattu. West of this island it swings northwest to a point south of Stalemate Bank, where it trends westward. Its upper limit is the edge of the Insular Shelf

or of the Ridge Shelf. East of Agattu its lower limit is the northern and inner edge of the Aleutian Bench, ranging in depth from 1600 to 2400 fathoms. West of Agattu the south slope which does so in greatly subdued form. Except for Etienne and Agattu, their southeast sides are straighter and steeper than their northwest sides, and most of them have rounded or box-

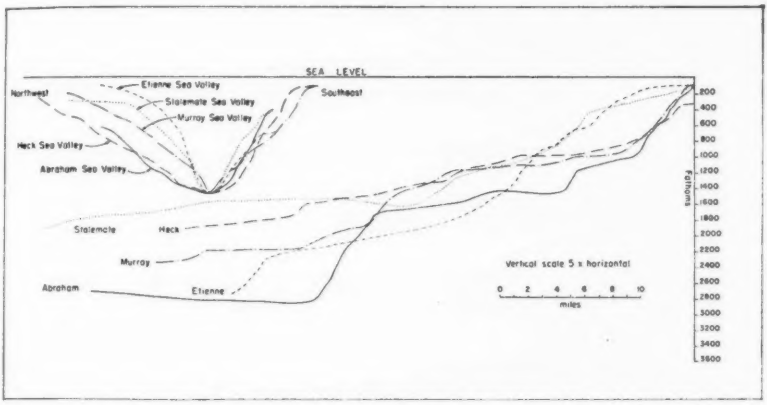

FIGURE 10.—CROSS-SECTIONAL AND LONGITUDINAL PROFILES OF FIVE SEA VALLEYS On South Insular Slope. Refer to Figure 3 for locations.

merges at an indefinite depth with the north side of the Aleutian Trench.

The average regional gradient of the South Insular Slope is 400 feet per mile south of Kiska and Amchatka and 450 feet south of Agattu. The gradient is not uniform. South of Kiska and Amchatka the profile is convex, and south of Agattu it is steplike. Elsewhere the South Slope is so indented by sea valleys and canyons that a figure for the average regional gradient is meaningless.

Large sea valleys cut the South Insular Slope; six are particularly prominent: Murray (Gibson and Nichols, 1953, Pl. 1), Heck (Pl. 1), Agattu, Abraham, Etienne, and Stalemate (Pl. 1). Abraham is more than 30 miles long and 15 miles wide at its mouth, almost as large as Attu. Relief is more than 16,000 feet from the head of Abraham Sea Valley to its mouth and more than 7000 feet on the steep scarp on the southwest side of the valley near its head. This submarine relief dwarfs that of Attu's highest peak, which reaches an altitude of about 3100 feet.

The large sea valleys trend between north-northeast and east-northeast, and none traverses the Insular Shelf except Abraham like heads. Their bottoms profiles are steplike and contain closed depressions in some of the treads (Figs. 10, 11).

The South Insular Slope also contains two sequences of buttresslike headlands below the southwest edge of the Near Islands Platform, and many small submarine canyons, scarps, knobs, and swales. Many of the topographic features of the South Slope, both large and small, are linear and lie at an angle to the maximum gradient of the surrounding slope.

Origin of the Insular Slopes.—One of the principal conclusions of this paper is that the submarine topography of the Insular Slopes primarily reflects the structure of the Aleutian Ridge. Figure 12 shows the distribution of the inferred faults. Most of them have been drawn on the premise that fracture zones are indicated by topographic features that display one, or preferably several, of the following characteristics:

(1) Straight or slightly curving steep slopes and scarps of more or less uniform gradient

(2) Straight, or, in a few cases, smoothly curving canyons

(3) Canyons with one side steeper than the other

(4) Canyons with closed depressions or steep pitches along their axes

(5) Alignment of the feature at an angle to the maximum gradient of the surrounding slope.

Some support for dependence on the above criteria is suggested by the general fault pattern (Fig. 12) which shows an approximately right-angle relationship among the inferred faults. Such a pattern is more typically structural than erosional. Also, this pattern is very similar to that of faults on the Near Islands (Fig. 2) and has the same general fault trends. Additional support for these criteria is presented below in discussion of the origin of some of the principal topographic features of the Insular Slopes.

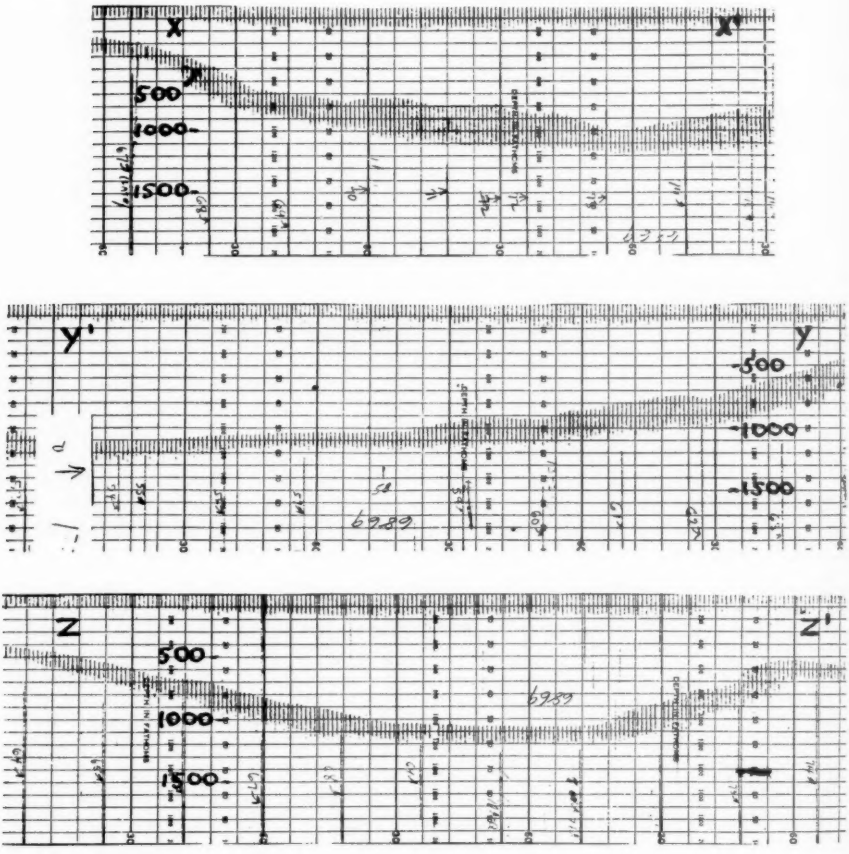

FIGURE 11.—FATHOGRAMS FROM HEAD OF ABRAHAM SEA VALLEY
Refer to Figure 4 for locations. Vertical scale approximately equals horizontal. Depths in fathoms.

The North Insular Slope is probably a great fault scarp marking a fundamental structural element of the Aleutian Ridge. The Slope is remarkably straight and its gradient of about 15° is much steeper than the 4° 17′ average gradient of typical continental slopes (Shepard, 1948, p. 187). Steep, straight or slightly curving submarine slopes are found in many places and have been interpreted as fault scarps. Typical examples are the scarps off the coast of southern California (Shepard and Emery,

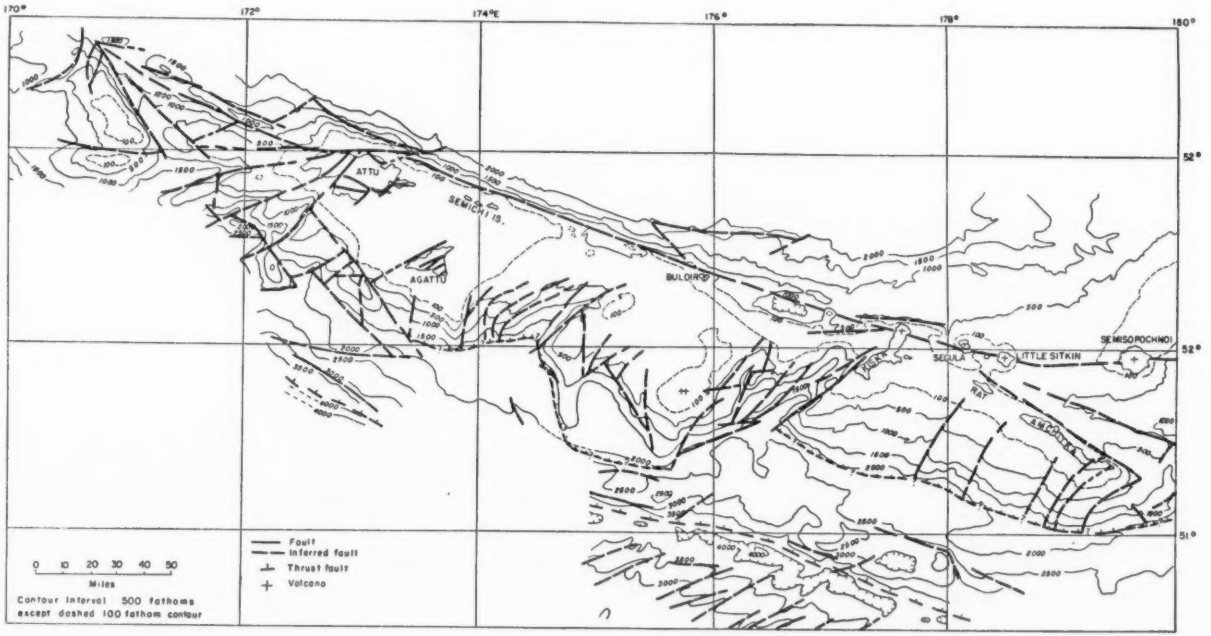

FIGURE 12.—DISTRIBUTION OF SUPPOSED FAULTS

Western part of Aleutian Ridge and Aleutian Trench. *See* Plate 1 and Gibson and Nichols, 1953, Pl. 1 for topographic details on which faults are based

1941, p. 9–49), the Sigsbee scarp in the Gulf of Mexico (Gealy, 1955, p. 214), and numerous scarps in the Caribbean region (Eardley, 1954, p. 744).

The alignment of volcanoes along the eastward extension of the North Insular Slope supports the conclusion that the latter represents a fracture zone. If the trend of the North Insular Slope is projected eastward along the slight curve of the Aleutian Ridge, it passes through, or very close to, the stratovolcanoes Buldir, Kiska, Segula, and Little Sitkin as well as the Buldir depression, the topography of which suggests a great caldera approximately 20 miles long, 10 miles wide, and 1 mile deep.

The dip of this great fault or fracture zone is believed to be northward. A dip to the south would require removal of a great volume of rock from the hanging wall, presumably by sliding and slumping. The apron at the base of the north slope is too small to contain a large enough volume of rock.

The linear ridges and troughs northwest of Attu parallel the North Insular Slope, suggesting relation to the same structural trend. The asymmetrical canyon off Holtz Bay is in line with a narrow depression on the Insular Shelf north of Attu and also with a steep slope along the north side of the submerged long ridge west of Attu. Apparently this alignment marks a fault that offset the North Slope opposite Holtz Bay. Steep north dips and contorted brecciated rocks along the north coast of Attu corroborate the presence of this fault.

The South Insular Slope has more complex topography than the North Slope, and its origin is not so evident. It may contain parts of an area that was near sea level in pre-middle Tertiary time, or parts of a pre-middle Tertiary slope to the Pacific Ocean floor, or both. Lower Tertiary sedimentary rocks on Attu contain leaves and a shallow-water pelecypod fauna; and conglomerate with well-rounded pebbles and cobbles occurs in the pre-Tertiary (?) spilitic rocks on Attu and Agattu. Prior to middle Tertiary orogeny, the area of the Near Islands was apparently near sea level and perhaps included some subaerial terrain. Presumably a slope to the Pacific Ocean floor also existed somewhere in the vicinity. Thus the South Insular Slope as a whole does not reflect uplift of the Aleutian Ridge from the floor of the Pacific during the middle Tertiary orogeny.

This pre-middle Tertiary slope or shallow terrain has been extensively warped and faulted, and its original topography has been destroyed except possibly for remnants preserved as the Aleutian Bench. The convex profile south of Kiska and Amchitka suggests a broad fold. The treads and risers south of Agattu and the buttresses and swales southwest of the Near Islands Platform perhaps indicate faults or folds parallel to the South Insular Slope.

Considerable evidence supports the conclusion that the great sea valleys and many of the submarine canyons of the South Insular Slope mark zones of faulting. Abraham, Etienne, and Stalemate sea valleys are aligned with known fault zones on western Attu, and the north-northeast to east-northeast trend of the sea valleys and canyons resembles the trend of major fault zones on the islands.

The positions of volcanoes near the points at which extensions of some sea valleys and canyons intersect the eastward extension of the North Slope (Fig. 12) further suggests that the sea valleys and some of the canyons are along fault zones. Buldir is near the extension of Heck Sea Valley; and Buldir Depression is in line with Coulee Canyon and the unnamed canyon on the northwest side of Coulee Canyon. Murray Sea Valley lines up fairly closely with Kiska volcano. The volcanic pile of Segula and Little Sitkin lies near the intersection of the extensions of Rat Island Canyon, Constantine Arm, and Oglalla Trough. Intersections of a great fracture like that represented by the North Insular Slope with prominent cross fractures would be logical places for volcanoes.

Shepard (1948, p. 229, 237) lists a sinuous trend, V-shaped cross section, absence of basin depressions, accordantly entering tributaries, and extension down, rather than along, submarine slopes as typical of continental slope submarine canyons. Kuenen (1950, p. 498) lists as evidence against such canyons being fault zones or fault troughs their "dendritic and sinuous pattern, the graded longitudinal profiles, the direction transverse to tectonic trends of the adjoining coast (California), the evidence linking with major aggrading rivers on adjoining land, the restricted time of formation

(late Pliocene to late Glacial), the great number off the east coast of the United States that is tectonically inactive." He joins Shepard and other students of submarine canyons in concluding that most submarine canyons on continental slopes result from erosion, although there are differences of opinion on the mechanism of the erosion (Daly, 1942, p. 136; Shepard, 1948, p. 207–249; 1951; 1952; Kuenen, 1950, p. 485–526; 1953, p. 1299–1304; Crowell, 1952).

The sea valleys and submarine canyons of the South Insular Slope differ considerably from typical continental slope canyons off the coast of southern California (Shepard and Emery, 1941) or off the coast of eastern United States (Veatch and Smith, 1939). The western Aleutian sea valleys and canyons, except for Bowie, lack a dendritic pattern, are straight or only slightly sinuous, have ungraded floors with closed basins and steep pitches, conform to structural trends on the islands, have no link with major aggrading streams, were probably formed mainly during the middle Tertiary orogeny, and are in a tectonically active area. In addition, they have marked asymmetry in their cross-section and lie at an angle to the maximum gradient of the surrounding slope. A tectonic origin explains these characteristics more readily than an erosional one.

Aleutian Bench

The Aleutian Bench is a prominent step in the general slope from the Aleutian Islands to the Aleutian Trench. It was first described by Murray (1945, p. 775) in the eastern Aleutians. Detailed hydrographic surveying is insufficient to establish whether the Aleutian Bench exists also in the central Aleutians. Within the limits of the Near Islands and the Rat Islands, the Aleutian Bench extends from south of Amchitka to the mouth of Heck Sea Valley and lies between the 1600- to 2400-fathom contours and the 3000-fathom contour. South of Agattu, remnants of the Bench may be represented by the treads in the irregular gradient of the South Insular Slope.

The Aleutian Bench consists of two topographic units. The inside of the Bench, opposite the mouths of Murray and Heck sea valleys and along the base of the convex slope south of Kiska and Amchitka, looks like a zone of sedimentation and shallow erosion perhaps by turbidity currents and landsliding. The outside of the Bench next to the Aleutian Trench has considerably more relief than the inside and contains cones and knolls, crude elliptical depressions, and broad valleys.

Conclusions regarding the origin of the Aleutian Bench must be extremely speculative. Sounding traverses in this area are 4 to 8 miles apart; there are no direct ties to the geology of the islands; and the Bench lacks marked linear topographic features that fit the criteria for structural interpretation.

Heck and Murray sea valleys and Rat Island Canyon apparently end at the inner edge of the Bench, suggesting that the faults represented by these valleys and canyons also stop at the inner edge. The Bench appears to mark a structural discontinuity, perhaps a thrust zone in which the faults represented by the sea valleys and canyons are restricted to the upper plate.

The topography of the outside edge of the Bench is puzzling. Kiska Knoll, James Knoll, and Thurmond Sea Mount look like volcanoes, and the general topography has a volcanic aspect. This very tenuous evidence hints that the Aleutian Bench may be a foundered sliver of volcanic terrain existing in the area during early Tertiary or pre-Tertiary time.

Aleutian Trench

The Aleutian Trench is an arcuate submarine trough extending about 1000 miles along the south side of the Aleutian Ridge from approximately 170° E. long. to south of Unimak Island. There the Trench swings south to pass outside Kodiak Island and ends off the mouth of Yakutat Bay in the Gulf of Alaska. Murray (1945) made the pioneer detailed study of the topography of the Trench, and Menard and Dietz (1951) added many details from the Gulf of Alaska area.

The comparatively small area of the Trench shown on Plate 1 and on the Rat Islands map (Gibson and Nichols, 1953, Pl. 1) clearly shows its asymmetry, a steep north side and a less steep south side. This asymmetrical cross sec-

tion and the arcuate longitudinal shape characterize the Trench for almost its entire length.

The north side of the Trench is indented by small V-shaped canyons, some of which traverse the slope at an angle. The pattern of the south side suggests an echelon folding or faulting. The floor is narrow, almost flat, and abruptly terminates the steep gradient and small canyons of the north side as well as the en echelon pattern of the south side. Perhaps the flat floor indicates a sedimentary fill. Some of Murray's (1945, p. 766–772) profiles further east show that the floor of the Trench has considerable relief and may not have a sedimentary fill everywhere.

The asymmetrical cross section and the arcuate longitudinal shape of the Trench are consistent with the hypothesis summarized by Umbgrove (1947, p. 148–214) that a deep trough along the convex side of an island arc marks the intersection of a thrust or "potential zone of shear" with the earth's surface. Benioff (1954) presents seismic evidence suggesting that great reverse faults dip beneath island arcs and reach the earth's surface in the vicinity of the oceanic trenches along the arcs. He (1954, p. 391) postulates a fault zone with a dip of about 28° N. under the Aleutian arc. If this hypothesis is accepted for the Aleutian Trench, the steep north side is presumably an expression of a drag fold formed along the nose of the upper plate as it moved southward over the Pacific Ocean floor. The en echelon pattern of the south side may indicate crumpling of the Pacific Ocean floor under stress from the advancing upper plate, the Aleutian Ridge.

Role of Submarine Erosion and Deposition

Although the submarine topography of the western part of the Aleutian Ridge is considered primarily tectonic, this does not imply that present topographic contours are also necessarily structure contours. Submarine slumping and landsliding, erosion and deposition by turbidity currents, and deposition of glacial and other material swept from the Crest of the Aleutian Ridge have all left their mark on the submarine topography.

Prominent scarps such as the North Insular Slope, the scarp northeast of Stalemate Bank, and the steep straight sides of many of the sea valleys have slopes that exceed 25° in few places. In contrast, dips of fault zones on the islands are 45° to vertical. The fathograms (Fig. 11) show the low gradient of the comparatively steep southwest side of Abraham Sea Valley, the hummocky profile typical of some of the scarps, and the flat floors typical of some of the sea valleys.

Originally steep fault scarps have probably been cut back by landsliding and slumping. Such processes might account for the hummocky profile of some of the scarps as well as the boxlike heads of some of the sea valleys. Shepard (1951) shows that mass movements are modifying the heads of canyons in the La Jolla, California, area. The flat floors of the sea valleys may be fill from landslides and slumps. The apron along the foot of the North Insular Slope looks like a slump or landslide deposit.

Turbidity currents also may have modified the topography. During glaciation much silt, clay, and coarser material must have been brought to the edge of the Insular Shelf, particularly by glaciers pouring down the main glaciated valleys of Attu such as Abraham and Etienne. Concentration of volcanic ash from eruptions would also make suitable raw material for turbidity currents. Such currents may have occasionally flushed unconsolidated material from the sea valleys and perhaps have cut V-shaped notches, such as those near the heads of Stalemate, Etienne, and Agattu sea valleys and also in the rises between steps in Murray and Heck sea valleys. The two shallow sinuous canyons winding across the Aleutian Bench from the mouth of Murray Sea Valley, other small sinuous canyons on the South Insular Slope, and some of the canyons on the north side of the Aleutian Trench may also be the product of turbidity-current erosion.

Structure of the Western Part of the Aleutian Ridge

This concluding section attempts to assemble the various structural elements of the Aleutian Ridge suggested by the submarine topography into a unified hypothesis portraying the overall structure of the Ridge.

The principal structural elements are: (1) the great fault along the north side of the Ridge represented by the North Insular Slope; (2) the faulted and warped South Insular Slope; (3) a possible thrust along the inside edge of the Aleutian Bench; (4) a thrust or zone of potential shear that dips beneath the Aleutian Ridge and is delineated by the Aleutian Trench.

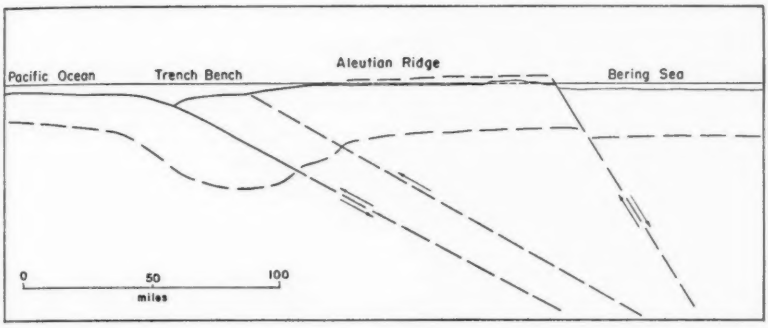

FIGURE 13.—DIAGRAMMATIC CROSS SECTION OF WESTERN PART OF THE ALEUTIAN RIDGE AND ALEUTIAN TRENCH

It is assumed that the Aleutian Trench is accompanied by a negative gravity anomaly such as those present with other oceanic trenches along island arcs and that these anomalies reflect a downbuckle of the earth's crust (Vening Meinesz, 1948, p. 25–90; 1954, p. 144–154). Some of the assumed anomaly may result from a thick fill of unconsolidated sediments in the Trench as Ewing and Worzel (1954, p. 171) postulate for the West Indies.

The Aleutian Ridge is pictured (Fig. 13) as an asymmetrical wedge bounded by a zone of northward-dipping normal faults on the north and by a zone of northward-dipping reverse faults on the south. The movement of this wedge southward caused a downbuckle and thickening of the earth's crust near the Trench. The sea valleys, submarine canyons, and other linear topographic features of the South Insular Slope perhaps mark zones of faulting, both normal and transverse, resulting from differential movement of blocks within the overriding wedge. The marine, subaerial, and glacial erosion that produced the Crest of the Aleutian Ridge reflects upward movement and arching of the wedge as it was thrust southward.

Stalemate and Heck sea valleys, Constantine Arm, Oglalla Trough, and the eastern part of the North Insular Slope end at the middle Tertiary (?) Ridge Shelf and are probably older than the latter. The correlation between inferred submarine faults and known faults on the islands indicates that the wedge probably began to form during the middle Tertiary orogeny preceding formation of the Ridge Shelf. Faulting and warping of the latter shows that deformation continued into the late Tertiary. As the Insular Shelf abruptly truncates the Insular Slopes and lacks apparent fault scarps, the deformation responsible for the Insular Slopes and their topography must have largely ceased by the last major Pleistocene glaciation at the latest. Frequent earthquakes and active volcanoes testify, however, that the Aleutian arc is still a restless part of the earth's crust.

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PETROLOGY AND WEATHERING OF THE IRON DIKE, BOULDER AND LARIMER COUNTIES, COLORADO

BY ERNEST E. WAHLSTROM

ABSTRACT

The Iron Dike in Boulder and Larimer counties, Colorado, is magnetite-rich diabase and can be traced diagonally across the central part of the Front Range of Colorado for nearly 38 miles. The fabric and mineralogy indicate emplacement by a single, rapid eruption originating in a deep, laterally wide-spread source. The magma was partly crystallized by the time it reached the level of the present surface exposures. Intrusive pressure continued during consolidation and produced fracturing and dislocation of previously formed crystals and promoted deuteric alteration by providing channels for concentration and movement of late residual solutions. The dike is chemically unlike and predates probably all the intrusive and extrusive rocks in the normal Cenozoic igneous sequence of the central Front Range.

Weathering of the dike near Sugarloaf, Colorado, caused extensive oxidation and hydration of iron-bearing minerals, especially pyroxene and chlorite, and downward transportation and deposition of colloidal limonitic material. Leaching has removed considerable calcium and magnesium from the upper part of the weathered zone. Crystallized clay minerals are notably absent.

Comparison of the weathering of the diabase and Precambrian granodiorite into which the dike was intruded indicates approximately the same amount of decomposition as measured by changes in the ferromagnesian minerals in both rocks. Deeper penetration of weathering effects in the granodiorite as compared to the diabase is attributed to a higher permeability in the granodiorite.

CONTENTS

TEXT

	Page
Introduction	148
Acknowledgments	149
Petrography	149
Fabric	149
Mineralogy	149
Chemical composition	151
Sequence of crystallization	151
Contact effects	152
Structural and magmatic relationships	152
Weathering changes in the Iron Dike	154
General statement	154
Physical environment and soil development	156
Gross chemical changes by weathering	156
Mineral reconstitution by weathering	159
Physical changes accompanying weathering	161
Summary of conclusions	162
References cited	162

ILLUSTRATIONS

Figure	Page
1. Index map	148
2. Paragenetic diagram	152
3. Larsen triangular plot	153
4. Cross section of Iron Dike near Sugarloaf	154
5. Weight per cent variation diagram for weathered diabase	156
6. Molecular per cent variation diagram for weathered diabase	157
7. Triangular diagram for FeO-Fe_2O_3-H_2O ratios	158
8. Triangular diagram for $(CaO + MgO)$-FeO-H_2O ratios	159
9. Cumulative diagram	162

Plate	Following page
1. Iron Dike exposures	150
2. Textural gradations	
3. Photomicrographs of diabase	
4. Photomicrographs of weathered diabase	

TABLES

Table	Page
1. Analyses of pyroxene	150
2. Optical properties of pyroxene	150
3. Analyses of fresh diabase	151
4. Analyses of fresh and weathered diabase, Sugarloaf exposure	155
5. Magnetite in fresh and weathered diabase, Sugarloaf exposure	160
6. Analyses of fresh and weathered granodiorite	160

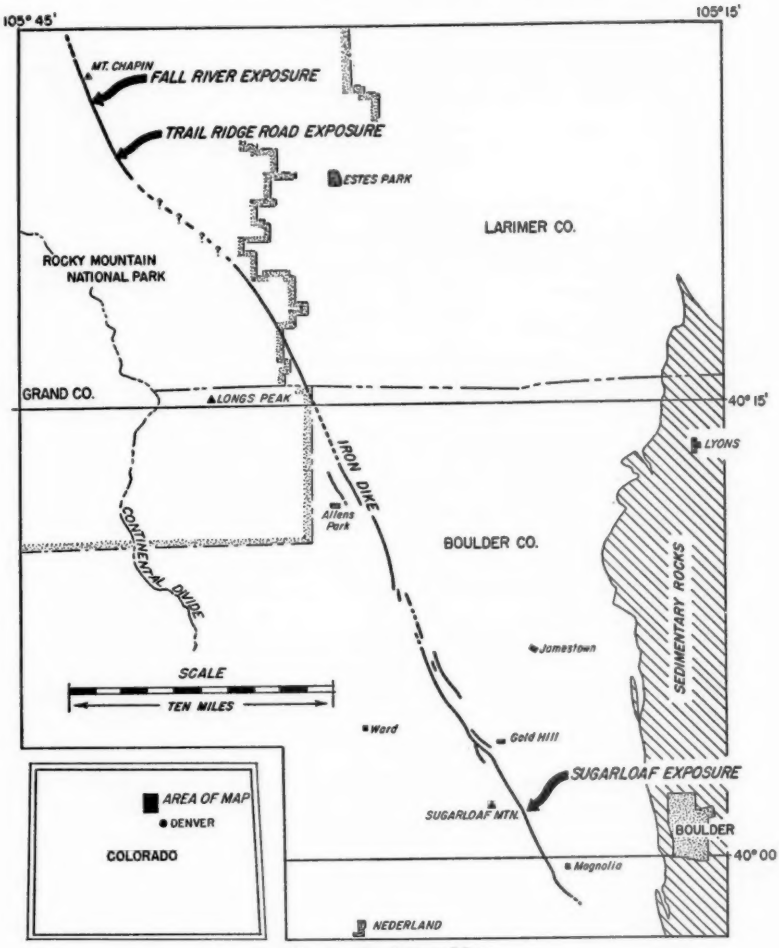


FIGURE 1.—INDEX MAP
Showing location of Iron Dike exposures

INTRODUCTION

The Iron Dike, an iron-rich diabase, can be traced, with interruptions, for nearly 38 miles diagonally northwest across the Front Range of Colorado from a point near the small mining town of Magnolia, west of Boulder, Colorado, through Mount Chapin in Rocky Mountain National Park (Fig. 1). Because it is more resistant to weathering and erosion than the granite, granodiorite, schist, and gneiss constituting its wall rocks, it forms a low, rounded ridge standing a few feet to as much as 100 feet above the land surface (Pl. 1, fig. 1).

The dike, named by early miners who noted the dark color, high specific gravity, high content of iron-rich minerals, and limonite-stained weathering products, strikes about N. 30° W. and dips 60°–80° SW. in most exposures. Characteristically it is 50–100 feet wide, but in some places is only a few feet or few inches wide.

There are remarkably few apophysal or subsidiary intrusions, and the composition of the

dike is unlike other mafic rocks in or near the Front Range in the vicinity of the dike.

Crawford (1908) mapped and described the petrography of the dike in the Sugarloaf District. Boos and Boos (1933) briefly describe its composition and distribution. Lovering and Goddard (1938; 1950) attempted to place the dike in its relative position in the Laramide igneous sequence of the Front Range.

In the present study special attention is given to the petrography and petrogenesis of the dike and the chemical and mineralogical changes that have resulted from weathering.

Acknowledgments

The writer gratefully acknowledges a generous grant for the purchase of special equipment by the Council on Research and Creative Work of the University of Colorado. The writer also acknowledges the assistance of William Braddock, Russell Honea, David Ramaley, and O. W. Tollefson, graduate students, who worked on special aspects of the Iron Dike. L. A. Warner and T. R. Walker of the Department of Geology, University of Colorado, read the manuscript and made valuable comments.

Petrography

Fabric

The fabric of the Iron Dike varies with distance from the wall rock, the amount of movement in the dike during solidification, and the intensity and distribution of deuteric alteration. Near and at the contacts rapid chilling produced a dense to partly glassy rock containing scattered phenocrysts of plagioclase and pyroxene embedded in a groundmass of minute crystals of pyroxene, feldspar, and magnetite. The dike shows an increase in grain size from the chilled contact toward the center (Pl. 2), where tabular grains of plagioclase and pyroxene appear in subhedral to anhedral crystals averaging 1–6 mm long. Both pyroxene and plagioclase tend to form clusters. Where the two minerals are in contact, anhedral pyroxene commonly fills interstices among plagioclase plates in an ophitic texture.

Movement in the dike in the intermediate and late stages of solidification locally produced crushing of plagioclase and pyroxene. Many of the crushed grains are healed with the same substances that formed the original crystals, an indication of continued growth after deformation. Curved cleavage planes, bent twin surfaces, and many splotchy plagioclase grains attest to the vigor and timing of the movement.

The long axes of the pyroxene and plagioclase crystals show a preferred orientation on contoured equal-area plots. Parallelism of orientation, well developed within individual clusters is less evident when all axes are plotted regardless of whether the crystals are in clusters or not. Preferred orientation is more pronounced in the contact zone than in the center. If it is assumed that the long dimensions of the crystals parallel the direction of movement, the magma moved upward along the dip as a sheet rather than upward and laterally from one or more centers of intrusion.

Mineralogy

Average modes by the point-counter method of chemically analyzed samples taken at the center of the dike from exposures at Sugarloaf, Trail Ridge, and Fall River canyon are as follows (volume per cent): plagioclase (labradorite), 49.8; pyroxene (augite), 38.4; titaniferous magnetite, 8.5; quartz, 2.3; orthoclase, 1.0. Samples from other exposures yielded similar results and indicate uniformity along the length of the dike. The mode and textures are characteristic of a magnetite-rich quartz-bearing gabbro, which, by decrease in grain size toward the walls, grades into a diabase or a glassy rock of the same composition.

The plagioclase as determined by immersion methods and the universal stage is complexly twinned labradorite (An_{55-60}) and is only slightly zoned. Comparison of plagioclase from the contact zone with larger crystals from the center indicates a slightly higher anorthite content in the crystals next to the contact. Many shattered crystals are cemented by more sodic feldspar or, in a few crystals, by a chloritic material. Numerous tiny flakes of sericite are present in many of the plagioclase grains. Stain

tests indicate that the mica is soda-rich and approximates paragonite.

The pyroxene forms subhedrons in single crystals and clusters or anhedrons interstitial to tabular plagioclase. Twinning on {100} is common. In some clusters feathery, wedge-shaped individuals are radially distributed. A poorly developed hour-glass structure is seen in some grains; skeletal crystals or separate plates of magnetite locally penetrate the crystals to produce a sagenitic structure.

Table 1.—Analyses of Pyroxene

	8	52	A
SiO_2	50.13	49.23	47.32
TiO_2	0.24	0.35	.....
Al_2O_3	7.05	5.73	6.37
Fe_2O_3	1.90	1.02	2.56
FeO	14.30	14.70	14.40
MnO	0.32	0.30	.....
MgO	11.16	11.90	13.43
CaO	14.43	14.89	16.08
Na_2O	0.76	0.92	.....
K_2O	0.33	0.66	.....
P_2O_5	0.09	0.19	.....
	100.71	99.89	100.16

8. Pyroxene from Sample No. 8. Center of dike, Sugarloaf exposure; analyst: E. E. Wahlstrom

52. Pyroxene from Sample No. 52. Center of dike, Trail Ridge exposure; analyst: E. E. Wahlstrom

A. Pyroxene from dike, Sugarloaf exposure; analyst: L. G. Eakins (Clarke, 1904, p. 187)

Exsolution lamellae of the type described by Poldervaart and Hess (1951, p. 481) were not observed in the pyroxene. However, chlorite, replacing the pyroxene, has revealed differences in the initial composition of various portions of the crystals by emphasizing the hour-glass structure and penetrating the crystals along closely spaced planes parallel to {001}. The chlorite probably was localized by submicroscopic exsolution lamellae differing in composition from the host crystal. The amount of chloritic alteration varies (Pl. 3) but is especially concentrated in the central portions of the dike where movement locally fractured the rock and provided channels. All gradations between delicately selective replacement and gross, nonselective replacement of entire grains or clusters may be seen within a single thin section.

Table 2.—Optical Properties of Pyroxene from Iron Dike at Sugarloaf

n	Orientation	
X = 1.688 ± .001		Biaxial positive
Y = 1.703 ± .001	= b	2V = 48°
Z = 1.713 ± .001	Ac = 42°	Birefringence = .025

With the use of bromoform and a Frantz magnetic separator pyroxene was separated from samples collected at Sugarloaf and Trail Ridge and was chemically analyzed. Microscopic examination of the powder used for analysis revealed small amounts of included magnetite and chlorite, which could not be removed. The analyses of the pyroxene and an incomplete analysis by Eakins of pyroxene from Sugarloaf are given in Table 1. Table 2 lists the optical properties of the pyroxene from the Sugarloaf exposure as determined by the oil-immersion method using a yellow light source. The analyses calculated into molecular percentages of CaO, MgO, and FeO yield points that fall within the augite field as specified by Poldervaart and Hess (1951, p. 474).

Titaniferous magnetite appears in equant

Plate 1.—IRON DIKE EXPOSURES

Figure 1.—Road Cut East of Sugarloaf, Looking Southeast
See Figure 4 for geologic cross section

Figure 2.—Spheroidal Weathering
Road cut east of Sugarloaf

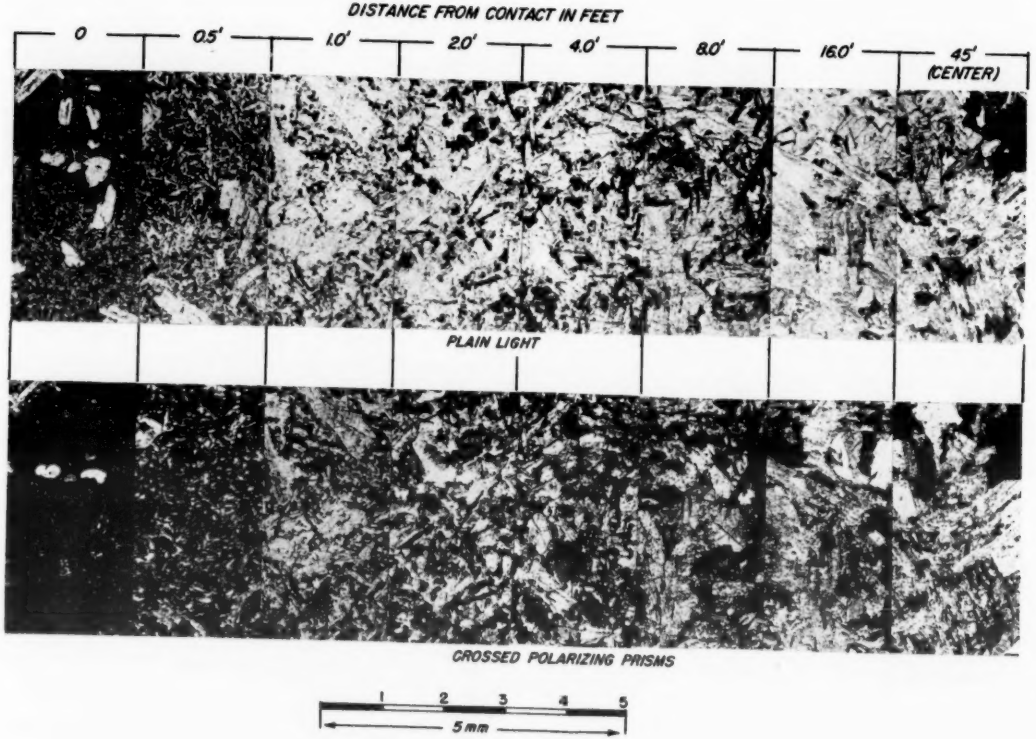
Plate 2.—TEXTURAL GRADATIONS

Photomicrographs of Thin Sections of Rock Samples Collected at Indicated Intervals from Contact of Dike toward Its Center

FIGURE 1

FIGURE 2

IRON DIKE EXPOSURES



TEXTURAL GRADATIONS

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PHOTOMICROGRAPHS OF DIABASE

FIGURE 2

FIGURE 1

PHOTOMICROGRAPHS OF WEATHERED DIABASE

crystals, plates, skeletal dendritic crystals, and in veins filling fractures in both pyroxene and plagioclase (Pl. 3). Growth during all stages of crystallization of the magma and intergrowth with other minerals has developed complex patterns in the magnetite in individual crystals and in multimineralic aggregates. A small amount of late magnetite replaced the pyroxene.

Quartz is a late mineral, probably deuteric. It fills interstices as scarce equigranular aggregates, replaces the plagioclase in myrmekitic intergrowths, and forms graphic intergrowths with orthoclase. Quartz is concentrated in the portions of the dike most fractured by late movements and is commonly accompanied by extensive chloritization of the pyroxene and sericitization of the plagioclase.

Scattered prisms of apatite in pyroxene-magnetite aggregates and isolated plates of biotite and narrow uralite rims around pyroxene are present in some samples.

Chemical Composition

Table 3 lists the chemical analyses of four samples from various exposures of the Iron Dike, an average of the four analyses, and the chemical composition of average diabase (Daly, 1933). A high content of Fe_2O_3 in sample 40 is explained by the fact that this sample is partly weathered and limonite-stained. Especially noteworthy is the similarity of the analyses despite considerable distances between points of collection (32 miles between samples 8 and 35) and the similarity between samples collected from the center and contact zone at the Trail Ridge exposure. Deuteric chloritization of the pyroxene in samples from the Trail Ridge and Fall River exposures is greater than in the Sugarloaf sample, an indication that chloritization has not caused appreciable changes in the bulk chemical composition of the rock.

TABLE 3.—ANALYSES OF FRESH DIABASE

(Analyst: E. E. Wahlstrom)

	8	35	40	52	Average	Averag Diabase (Daly, 1933)
SiO_2	48.75	49.05	48.11	48.71	48.65	50.48
TiO_2	0.74	0.73	0.73	0.74	0.73	1.45
Al_2O_3	17.64	16.32	16.35	16.47	16.69	15.34
Fe_2O_3	1.51	2.78	5.88	2.90	3.27	3.84
FeO	11.64	11.49	9.03	12.01	11.04	7.78
MnO	0.15	0.14	0.10	0.19	0.14	0.20
MgO	5.02	4.56	4.27	4.47	4.58	5.79
CaO	8.34	8.92	8.55	7.85	8.41	8.94
Na_2O	3.80	4.11	5.39	4.60	4.47	3.07
K_2O	1.32	1.53	1.13	1.17	1.29	0.97
P_2O_5	0.41	0.41	0.44	0.53	0.45	0.25
H_2O+	0.42	0.50	0.22	0.63	0.44	1.89
H_2O-	0.18	0.18	0.22	0.15	0.18	
CO_2	Tr	Tr	Tr	None	....	....
	99.92	100.72	100.42	100.42	100.34	100.00

8. Center of dike, Sugarloaf exposure
35. Center of dike, Fall River Road exposure
40. Dike at contact with granite, Trail Ridge Road exposure
52. Center of dike, Trail Ridge Road exposure

Sequence of Crystallization

Figure 2 illustrates the paragenesis of the Iron Dike. At the level of observation in the surface exposures of the dike plagioclase, pyro-

PLATE 3.—PHOTOMICROGRAPHS OF DIABASE

FIGURE 1.—DENDRITIC, SKELETAL CRYSTALS OF MAGNETITE (BLACK) IN PLAGIOCLASE AND PYROXENE
FIGURE 2.—IRREGULAR VEINLETS OF MAGNETITE (BLACK) IN PLAGIOCLASE (CLEAR)
FIGURE 3.—CHLORITE (C) REPLACING PYROXENE (P)
Clear mineral of low relief is labradorite (L); plain light
FIGURE 4.—SAME AS FIGURE 3
Crossed polarizing prisms

PLATE 4.—PHOTOMICROGRAPHS OF WEATHERED DIABASE

FIGURE 1.—SELECTIVE REPLACEMENT OF CHLORITE AND PYROXENE BY LIMONITIC MATERIAL (BLACK)
Plain light
FIGURE 2.—GROSS REPLACEMENT OF PYROXENE BY LIMONITIC MATERIAL (BLACK)
Note shrinkage cracks in limonite. Clear areas are labradorite. Plain light

xene, and magnetite were in the magma in small amounts at the time of intrusion. Resorption is evident in rounded and embayed pyroxene and plagioclase crystals in the contact zone, but, almost immediately after intrusion, the crystals began to grow again. Continued, overlapping growth proceeded with maximum velocity near the contact and more slowly toward the interior. Movement continued in the partly liquid interior until the magma was completely frozen and resulted in complex fracturing and crushing of existing crystals and subsequent rehealing by continued crystallization.

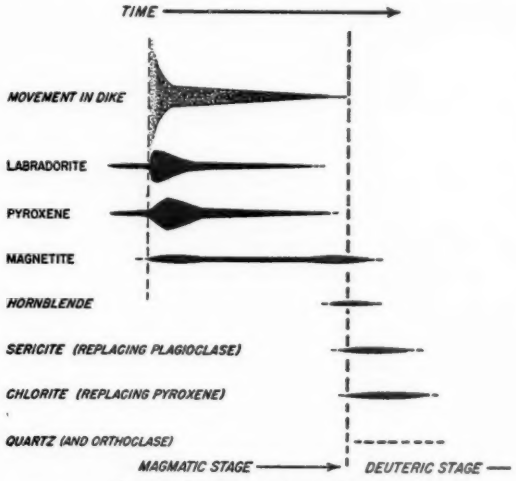


FIGURE 2.—PARAGENETIC DIAGRAM

Shows time of movement in dike and crystallization of magmatic and deuteric minerals as a function of time

Toward the end of the period of solidification residual deuteric solutions attacked the pyroxene and formed chlorite, introduced sericite into the plagioclase, and deposited quartz in interstices in myrmekitic intergrowths with the plagioclase and in micrographic intergrowths with potassium feldspar. Hornblende formed in small amounts at the expense of the pyroxene.

CONTACT EFFECTS

Boos and Boos (1933) attribute masses of magnetite containing rosettes of sillimanite and pleonaste (spinel) in the wall rocks at the Trail Ridge exposure to penetration of the "basic end products of the iron dike intrusion". The writer did not observe the rosettes. In most exposures dense or partly glassy dike rock makes a sharp contact with unmetamorphosed wall rocks. At the Sugarloaf, Trail Ridge, and Fall River exposures, late deuteric solutions have invaded granite and gneiss wall rocks and have formed sericite at the expense of the plagioclase.

Scattered, small inclusions of the wall rocks are locally abundant in the dike. Some of the inclusions have been melted, especially those of granitic composition, and have been converted wholly or partly into a dense, glassy substance sericitized in the same manner and at the same time as the dike rock.

STRUCTURAL AND MAGMATIC RELATIONSHIPS

The Iron Dike filled a northwesterly trending fissure that parallels a series of persistent mineralized fault zones (locally called "breccia reefs") extending across the central part of the Front Range west of Boulder (Lovering and Goddard, 1950, p. 79). In Boulder County the dike is intersected and offset by numerous east-west faults, some of which have localized tungsten minerals and tellurides within the dike and its wall rocks. Later dikes (in the Jamestown district a hornblende diorite dike,

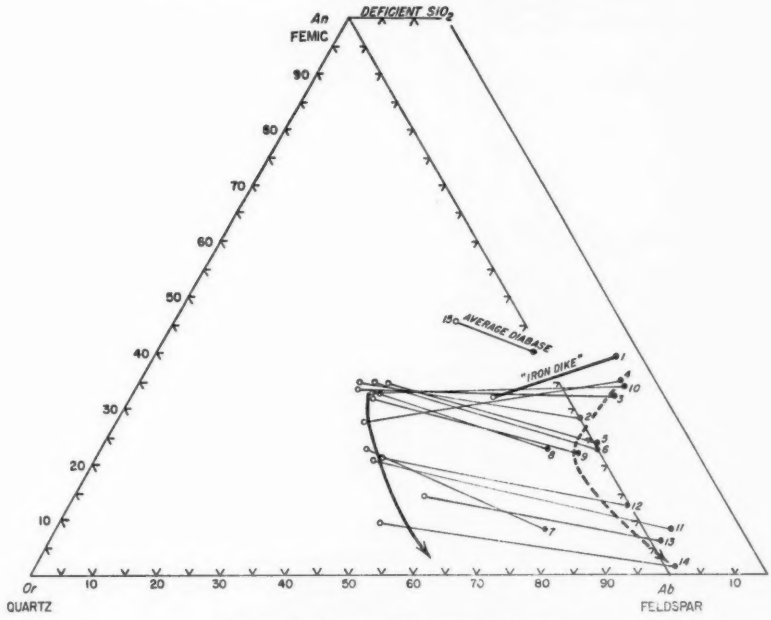


FIGURE 3.—LARSEN TRIANGULAR PLOT

Shows positions of Daly's average diabase and Iron Dike diabase relative to selected Cenozoic igneous rocks of central and northern part of Front Range, Colorado. Numbers identify analyses and calculated points for rocks. Open circles represent normative feldspars, and solid circles indicate quartz-feldspar-femic ratios. Arrows suggest differentiation trends for the normative feldspars (solid arrow) and quartz-feldspar-femic normative minerals (dashed arrow).

1. Average. Iron Dike.
2. Basalt. Table Mountain near Golden, Colorado (Emmons, Cross, Eldridge, 1896, p. 306). Analyst: W. F. Hillebrand.
3. Basalt. Table Mountain near Golden, Colorado (Emmons, Cross, Eldridge, 1896, p. 308). Analyst: W. F. Hillebrand.
4. Dolerite (diabase). Valmont dike, east of Boulder, Colorado (Emmons, Cross, Eldridge, 1896, p. 301). Analyst: L. G. Eakins.
5. Gabbro. Center of Caribou stock, Boulder County, Colorado (Jennings, 1913, p. 16). Analyst: E. P. Jennings.
6. Gabbro. Edge of Caribou stock, Boulder County, Colorado (Jennings, 1913, p. 16). Analyst: E. P. Jennings.
7. Andesite. Sugarloaf Mountain, Boulder County, Colorado (Hogarty, 1899, p. 181). Analyst: B. Hogarty.
8. Quartz monzonite. Caribou stock, Boulder County, Colorado (Washington, 1917, p. 471). Analyst: G. Steiger.
9. Quartz monzonite. Caribou stock, Boulder County, Colorado (Washington, 1917, p. 471). Analyst: G. Steiger.
10. Syenogabbro. Audubon-Albion stock, Boulder County, Colorado (Wahlstrom, 1940, p. 1806). Analyst: E. E. Wahlstrom.
11. Nephelite-bearing monzonite. Audubon-Albion stock, Boulder County, Colorado (Wahlstrom, 1940, p. 1806). Analyst: E. E. Wahlstrom.
12. Monzonite. Audubon-Albion stock, Bolder County, Colorado (Wahlstrom, 1940, p. 1806). Analyst: E. E. Wahlstrom.
13. Monzonite. Audubon-Albion stock, Boulder County, Colorado (Wahlstrom, 1940, p. 1806). Analyst: E. E. Wahlstrom.
14. Quartz-bearing monzonite. Audubon-Albion stock, Boulder County, Colorado (Wahlstrom, 1940, p. 1806). Analyst: E. E. Wahlstrom.
15. Average diabase (Daly, 1933, p. 18).

near Gold Hill a rhyolite porphyry dike, and near Sugarloaf a limburgite dike) cut the Iron Dike (Lovering and Goddard, 1938). Lovering and Goddard (1938, p. 48; 1950, p. 47) regard the intrusion of the Iron Dike as an early event

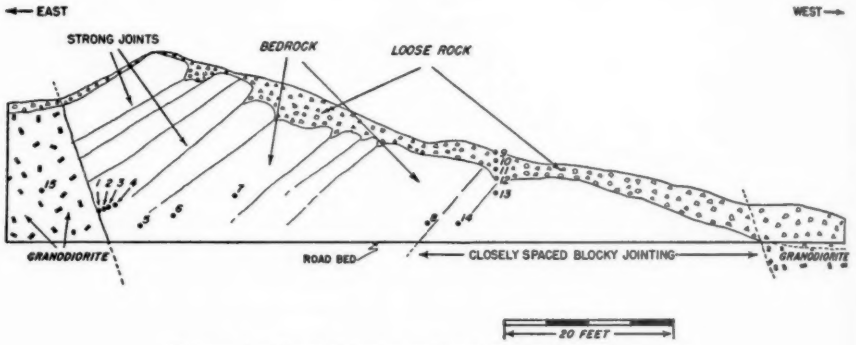


FIGURE 4.—CROSS SECTION OF IRON DIKE NEAR SUGARLOAF
Numbers indicate sample locations

in the igneous sequence that accompanied Laramide deformation and uplift of the Front Range. On the basis of chemical compositions they correlate the Iron Dike with basaltic lava sheets interbedded with sediments in the upper part of the Denver formation at Table Mountain near Golden, and with dolerite in a dike near Valmont. This correlation indicates a Paleocene age.

However, comparison of chemical analyses of the Iron Dike with analyses of the basalt at Table Mountain and the dolerite at Valmont emphasizes dissimilarities. Wahlstrom (1940) correlated the oldest intrusion in the composite Audubon-Albion stock in western Boulder County, a syenogabbro, with the Valmont dike and the Table Mountain flows. All these rocks have similar chemical compositions, characteristically high-potassium concentrations, and appear to have been derived from a widespread, deep source that after differentiation and assimilation provided most of the great variety of Cenozoic intrusive rocks in the Front Range area west of Boulder.

The chemical differences between the Iron Dike diabase and Cenozoic intrusive rocks in near-by portions of the Front Range are emphasized by use of a triangular plot (Fig. 3) (Larsen, 1938). Two points are plotted for each rock, one for normative feldspars (in silica-deficient rocks feldspathoids are calculated to feldspar), and the other for normative quartz, feldspar, and femic minerals. When nephelite appears in the norm it is multiplied by 1.845 to convert it to albite. If olivine is present, it is multiplied by 1.43 for forsterite and 1.29 for fayalite to produce pyroxene. The silica deficiency is the amount of silica added to nephelite and olivine to convert them to saturated minerals.

Figure 3 shows that the Iron Dike is not part of the early and middle Cenozoic igneous sequence of the central and northern parts of the Front Range. The point for normative feldspars for the Iron Dike does not fall near the curve indicating the trend of differentiation for normative feldspars of the Front Range rocks.

The conclusion that the chemical features of the other Cenozoic intrusive rocks of the central Front Range are lacking in the Iron Dike diabase implies that it was formed by a separate process and at a separate time, and is not part of the normal igneous sequence. Possibly the dike was intruded during or before the earliest manifestations of Laramide orogeny.

WEATHERING CHANGES IN THE IRON DIKE

General Statement

A chemical and mineralogical study was undertaken to determine changes resulting from weathering. Because the Iron Dike is in a mountainous region and is exposed over a considerable horizontal length, exposures are

TABLE 4.—ANALYSES OF FRESH AND WEATHERED DIABASE, SUGARLOAF EXPOSURE
(Analyst: E. E. Wahlstrom)

	9	10	11	12	13	14	8
SiO_2	50.09	48.48	47.24	47.63	48.20	47.92	48.75
TiO_2	0.70	0.70	0.70	0.70	0.74	0.72	0.74
Al_2O_3	18.87	17.95	16.78	19.62	16.46	16.15	17.64
Fe_2O_3	7.48	11.09	12.51	10.83	9.23	8.56	1.51
FeO	4.67	3.29	3.35	2.63	5.49	6.57	11.64
MnO	0.18	0.22	0.27	0.37	0.21	0.20	0.15
MgO	2.46	1.92	2.65	2.09	3.54	4.84	5.02
CaO	4.11	3.82	3.86	3.19	5.08	6.69	8.34
Na_2O	4.51	3.88	3.94	4.23	3.83	3.32	3.80
K_2O	2.48	1.96	1.85	1.80	2.06	1.42	1.32
P_2O_5	0.37	0.42	0.43	0.41	0.52	0.41	0.41
H_2O^+	3.87	4.56	4.49	4.79	2.97	2.66	0.42
H_2O^-	0.91	1.45	1.59	1.33	1.21	1.07	0.18
CO_2	None	Tr	Tr	None	None	Tr	Tr
	100.70	99.74	99.66	99.62	99.99	100.53	99.92
CALCULATED TO A WATER-FREE BASIS							
SiO_2	52.22	51.72	50.49	50.97	50.34	49.54	49.09
TiO_2	0.73	0.75	0.75	0.70	0.74	0.72	0.74
Al_2O_3	19.67	19.15	17.93	17.80	17.19	16.69	17.76
Fe_2O_3	7.80	11.83	13.36	14.78	9.63	8.84	1.52
FeO	4.87	3.51	3.58	2.81	6.21	6.79	11.72
MnO	0.19	0.23	0.29	0.40	0.21	0.20	0.15
MgO	2.56	2.05	2.83	2.24	3.69	5.00	5.05
CaO	4.28	4.08	4.12	3.41	5.31	6.91	8.40
Na_2O	4.70	4.14	4.21	4.52	4.01	3.43	3.83
K_2O	2.59	2.09	1.98	1.93	2.15	1.47	1.33
P_2O_5	0.39	0.45	0.46	0.44	0.52	0.41	0.41
	100.00	100.00	100.00	100.00	100.00	100.00	100.00

9–14. Weathered diabase

8. Fresh diabase from center of dike; Sugarloaf Road exposure. For location of samples *see* Figure 4.

found in a great variety of environments. A road cut about a quarter of a mile east of the town of Sugarloaf offered the best exposure for collection of samples grading from intensively weathered rock at the surface into fresh rock in the deeper portions. The Sugarloaf exposure has not been disturbed by recent stream erosion or glacial scouring and probably is more nearly in chemical equilibrium with its environment than are other exposures. For these reasons the weathering study was confined to the Sugarloaf exposure.

The dike at the Sugarloaf exposure is approximately 90 feet wide and dips about 70° SW. (Fig. 4). Three joint sets at right angles produce conspicuous joint blocks in some portions. The best-developed joint set is perpendicular to the walls and has the same strike as the dike. Weathering has reduced the surface exposures to an iron-stained rubble a few inches to a few feet thick. The wall rocks are gneissoid biotite-hornblende granodiorite of the Precambrian Boulder Creek batholith. The granodiorite is thoroughly weathered, and, although most of the original textures are preserved, the rock is decomposed so that it can be dug with ease. The granodiorite has been more pervasively weathered, and the effects of the weathering extend below the level of observation.

Physical Environment and Soil Development

The Sugarloaf exposure is at an elevation of 7650 feet. The adjacent area is well drained, and within a quarter of a mile to the west, north, and south of the road cut the land surface is relatively flat, except for the low ridge held up by the dike. To the east, the land surface slopes down into a dry, open, grass-covered valley.

The exposure is partly covered by an open stand of ponderosa pine and a sparse undergrowth of grasses, low shrubs, and flowering herbs. Many exposures of the dike are barren of vegetation and are covered by a brownish sandy aggregate containing scattered boulders. Black topsoil, where present, is very thin, and a sharply zoned soil profile is lacking.

The present climate is subhumid. Summers are dry with moderate temperatures, and the winters are long and cold. The Sugarloaf exposure of the dike is about 5 miles east of the easternmost limits of the valley glaciation that scoured the higher portions of the Front Range during pleistocene; the climate during late Pleistocene may have been periglacial in the Sugarloaf vicinity. However, the existing soil and rock debris covering the dike show no extensive frost or ice disturbance or other periglacial phenomena. The soil is a lithosol of the Underwood-Babb type as classified by the United States Department of Agriculture (Ableiter, 1938, p. 1126–1127).

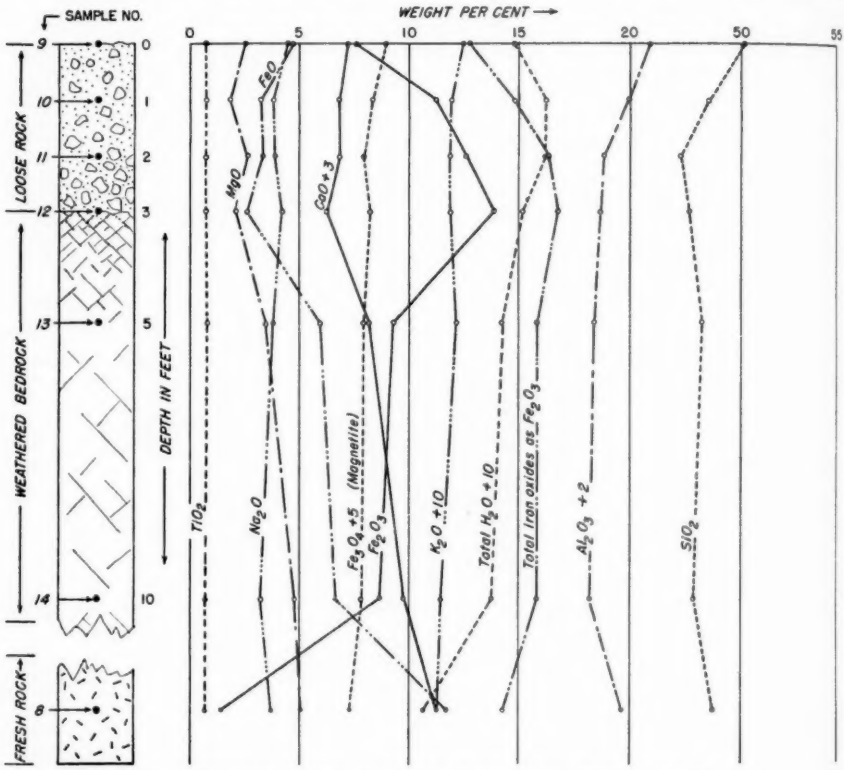


FIGURE 5.—WEIGHT PER CENT VARIATION DIAGRAM FOR WEATHERED DIABASE
Road cut east of Sugarloaf

Gross Chemical Changes by Weathering

Samples were analyzed to determine the gross chemical changes resulting from weather-

ing (Table 4). Microscopic examination of samples of loose material at and near the surface revealed small amounts of wind-blown quartz and microcline which came from the diabase has been weathered are the amount of water and the amount of ferric oxide formed at the expense of the ferrous oxide at various depths. The amount of ferric oxide in Sample

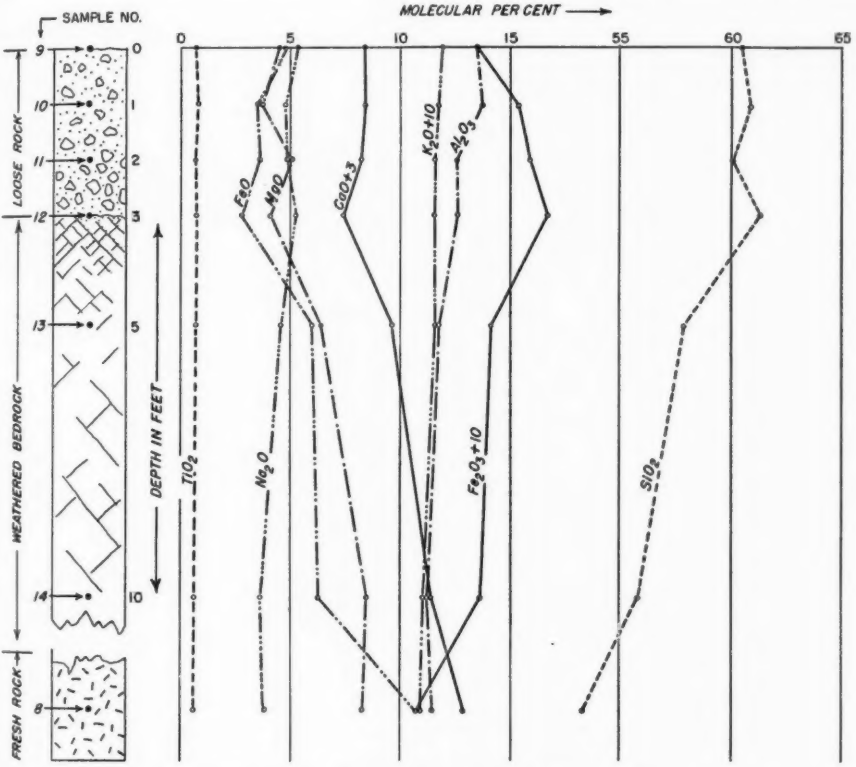


FIGURE 6.—MOLECULAR PER CENT VARIATION DIAGRAM FOR WEATHERED DIABASE
Road cut east of Sugarloaf

weathered granodiorite. This contamination was considered insufficient to affect the analyses significantly. The data in Table 4 were analyzed by the graphical devices shown in Figures 5, 6, 7, and 8.

Figure 5 shows the weight per cents of the oxides plotted against depth. Figure 6 shows molecular percentages on a water-free basis. Both diagrams emphasize the vertical changes that resulted from weathering. Variations in the amounts of water and iron oxides and the state of oxidation of the iron with depth are especially noteworthy and indicate that the most sensitive measures of the extent to which 14, from a depth of 10 feet, compared with ferric oxide in fresh diabase (Sample 8) collected 5 feet away, and the relative amounts of water in the two samples emphasize the relationship between penetration of the dike by ground water and the extent of oxidation of the iron. The ferrous oxide minimum and the ferric oxide maximum are at about the same depth (3 feet) at the base of the broken rock, but the total water content is at a maximum between 1 and 2 feet. The curve for total iron oxides shows a maximum at 3 feet and a minimum at the surface, indicating surface leaching, downward movement, and concentration of iron

oxides at the base of the loose rock. The total amount of iron oxides at a depth of 10 feet is greater than the total in adjacent exposures of fresh diabase, proof that ground waters transported iron downward to the maximum depth of weathering.

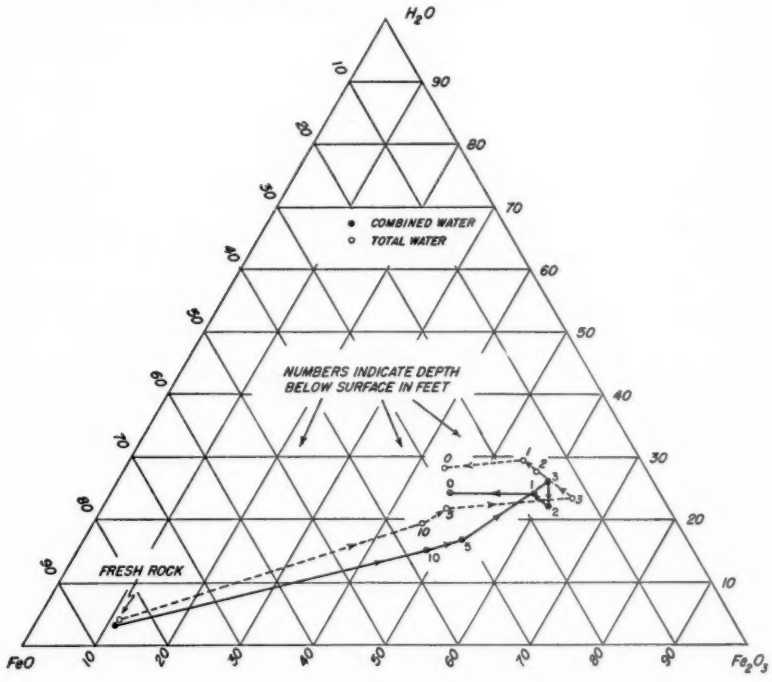


FIGURE 7.—TRIANGULAR DIAGRAM FOR FeO-Fe_2O_3-H_2O RATIOS

The relationships among ferrous oxide, ferric oxide, and water in the altered diabase are emphasized in Figure 7. The diagram shows variations of iron oxides with combined water, which is the amount of water driven off during chemical analysis at temperatures above 105°C, and total water, which includes both combined water and mechanically held water. Inasmuch as the amount of total water varies seasonally, combined water is a better measure of the extent of hydration. Most of the combined water is in limonite.

In Figure 7 the ferric oxide maximum and the ferrous oxide minimum are evident at a depth of between 1 and 3 feet. Between a depth of 1 foot and the surface ferrous oxide increases while the amount of combined water remains constant.

The increase in ferrous oxide from the base of the loose mantle toward the surface was unexpected. To test the idea that systematic mechanical concentration of residual magnetite at or near the surface of the weathered zone might account for the relatively high ferrous oxide concentration, magnetite was separated and weighed. The results are shown in Table 5 and as plotted on Figure 5 show insignificant variations in the magnetite contents of fresh and altered diabase compared to variations in the ferrous oxide contents, proving that variations in the ferrous oxide content of the weathered rocks are independent of the magnetite content.

Organic material in the upper few inches of

the mantle probably has been the agent in reducing ferric oxide to ferrous oxide. Organic material and drier conditions in the upper portion of the loose rock cover probably account for the high ferrous oxide concentration there.

Figures 5 and 6 indicate significant variations in lime and magnesia as well as in iron oxides and water. Lime and magnesia were leached in approximately the same ratios at various depths. Chemical analysis and thin sections show that calcium and magnesium carbonates are not present in the weathered rock—an indication that dissolved lime and magnesia were not retained in the rock to any extent. On the assumption that water and ferrous oxide are a measure of weathering, the sum of CaO + MgO was plotted against water and ferrous oxide (Fig. 8). Figure 8 shows a close correlation between the water maxima and the CaO + MgO minima and illustrates the progressive leaching of lime and magnesia during weathering.

Among the other oxides only silica and alumina show significant variations. Both enrich the upper portion of the loose mantle and are residual concentrations.

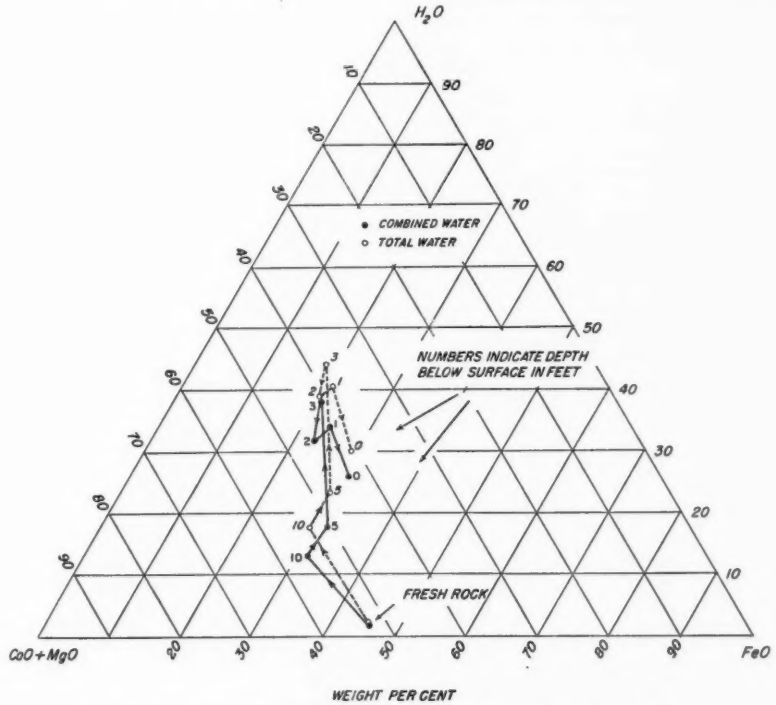


FIGURE 8.—TRIANGULAR DIAGRAM FOR (CaO + MgO)-FeO-H_2O RATIOS

To compare the weathering changes in the diabase with those in the granodiorite adjacent to the dike, fresh and weathered samples of granodiorite were analyzed (Table 6). As in the diabase, weathering resulted in removal of lime and magnesia and addition of water. Ferrous oxide in the fresh granodiorite was oxidized, and some iron was added by downward movement of ground waters.

Mineral Reconstitution by Weathering

The most abundant product of weathering is yellow to brown limonite which fills cracks,

coats mineral grains, and spreads pervasively through relatively unjointed rock. X-ray powder photographs of the limonite yield faint goethite lines and indicate that the material is essentially noncrystalline. Chemical tests reveal small amounts of amorphous alumina and silica. The maximum concentration of limonite is at and near the base of the loose rock (1–3 feet). Smaller amounts of limonite in the upper foot and fracture fillings of limonite at the maximum depth of penetration indicate leaching of iron compounds at and near the surface and downward transportation to the maximum depth of penetration by weathering solutions. The concentration of limonite at the base of the loose mantle indicates greater porosity here than in the bedrock.

TABLE 5.—MAGNETITE IN FRESH AND WEATHERED DIABASE, SUGARLOAF EXPOSURE*

Weight Per Cent

9	10	11	12	13	14	8
2.3	2.8	2.9	3.1	2.9	3.2	3.9
Water-Free						
2.3	2.9	3.0	3.3	3.1	3.4	4.1

* For location of samples *see* Figure 4.

Nearly all the limonite formed at the expense of augite or mixtures of augite and chlorite (Pl. 4). All degrees of alteration to limonite may be observed, and, depending on proximity to the surface exposure of the dike and the presence of fractures in the bedrock, the alteration grades from delicately selective staining to gross replacement affecting entire grains. Limonitic alteration of some augite crystals along closely spaced planes was probably localized by submicroscopic exsolution lamellae of a pyroxene richer in magnesium than the bulk of the crystal; in other crystals only the outer portions were altered, emphasizing an hour-glass structure. Chlorite in pyroxene was not converted to limonite until the associated augite was almost or completely destroyed. The limonite formed from augite and chlorite is impure and contains amorphous silica and aluminous materials.

The formation of limonite at the expense of augite and chlorite was accompanied by a volume decrease; many limonite pseudomorphs contain shrinkage cracks, some of which have localized veinlets of quartz.

TABLE 6.—ANALYSES OF WEATHERED AND FRESH GRANODIORITE

(Analyst: E. E. Wahlstrom)

	15	16
SiO_2	62.85	63.70
TiO_2	0.16	0.15
Al_2O_3	17.13	17.64
Fe_2O_3	4.31	1.67
FeO	0.60	1.80
MnO	0.12	0.20
MgO	1.65	2.60
CaO	1.72	3.83
Na_2O	4.04	4.06
K_2O	2.79	2.69
P_2O_5	0.36	0.30
H_2O+	3.02	0.76
H_2O-	0.88	0.37
CO_2	None	None
	99.63	99.77

15. Composite sample, weathered granodiorite, adjacent to diabase dike; Sugarloaf

16. Composite sample, fresh granodiorite, vicinity of diabase dike; Sugarloaf

Staining by limonitic material affects all the minerals in the intensively weathered upper portions of the exposure. Some of the calcic plagioclase is a faint yellow. Sericite replacing the plagioclase has been especially susceptible to absorption of limonitic stain and in many aggregates has been invaded in preference to the plagioclase.

A wavelike advance of limonite into fresh rock is evident near joints where rhythmic precipitation has produced alternating iron-rich and iron-poor layers parallel to the walls. In the deeper weathered zone six or eight layers of limonite concentration may occur within a fraction of an inch from the wall of a joint. Near the surface or near prominent joints limonitic stain has advanced to the centers of spheroidally weathered joint blocks (Pl. 1, fig. 2).

Part of the limonite stain was introduced

from solutions or suspensions occupying joints or cracks. In the walls of some of the joints a wave of limonitic stain has bypassed pyroxene grains. In other places limonite has selectively stained sericite plates, although there is no evidence of hydration or oxidation of adjacent iron-rich minerals.

The mechanism whereby the limonitic material entered the fresh rock and stained it is not apparent. X-ray tests of the limonite in the joints reveal its colloidal nature and suggest colloidal transportation.

Pervasive staining of the sericite in plagioclase probably resulted from local diffusion and absorption in the micaceous structure. Other minerals are stained internally or not at all, and only the exteriors of individual grains are coated. Presumably movement and deposition of the limonite was localized by original permeability or permeability induced by incipient weathering confined to grain boundaries.

Feldspar grains in the most intensively altered portion of the exposure show only slight cloudiness resulting from clay-mineral alteration. Grains containing abundant deuteric sericite show almost no clay mineral alteration; seemingly, the sericite has exerted a protective influence. Many otherwise unaltered grains of plagioclase in the loose mantle rock are rounded and embayed and have been etched and partly dissolved by ground water.

Repeated attempts were made to identify clay minerals, but, although large volumes of finely crushed, weathered rock were concentrated, only weak indications of crystalline clay minerals were found. Poorly developed differential thermal maxima and minima and weak, diffuse x-ray patterns indicate small amounts of a clay mineral of the kaolinite group, and illite, possibly formed by hydration of the sericite.

Several thin sections were made to compare weathering changes in the diabase with those in the granodiorite wall rock, which in fresh exposures consists of a medium- to coarse-grained aggregate of microcline, oligoclase, quartz, biotite, hornblende, and accessory minerals. Alterations in the granodiorite are almost identical with those resulting from recent weathering of the same type of rock on Flagstaff Mountain west of Boulder (Wahlstrom, 1948).

Weathering of the granodiorite in most exposures has resulted in extensive alteration of the plagioclase to clay minerals, although adjacent to the dike, where the plagioclase is extensively sericitized, clay alteration is almost absent. Quartz and microcline were not appreciably affected by the weathering, but the hornblende has been completely changed to limonite, and the biotite has been converted to bleached hydrobiotite.

A quantitative comparison of the intensities of weathering of the granodiorite and the diabase is difficult because of the different original mineralogy and textures. The diabase dike is more resistant to weathering and erosion. Iron staining and induced friability penetrate to greater depths in the granodiorite, but this is probably due to greater original permeability.

In both rocks the ferromagnesian minerals were attacked first. In the granodiorite biotite and hornblende were altered at the same time and to approximately the same degree as the pyroxene and chlorite in the diabase. Oligoclase in the granodiorite was converted to clay minerals more rapidly than was the labradorite in the diabase, but deuteric sericitization of the plagioclase in both rocks deterred weathering alteration. In both rocks original magnetite, potassium feldspar, and quartz were only slightly affected.

Physical Changes Accompanying Weathering

Hydration of iron oxide and removal of soluble constituents of the weathered diabase produced volume changes which could not be estimated. Gravitative movement of the loose rock in the mantle caused some downslope dislocation, but the attitudes, shapes, and locations of rock fragments indicate that most of the mantle is essentially residual and in place. Frost action has caused no extensive dislocation or redistribution of the loose material; the high permeability and excellent drainage prevented concentrations of ground water in the mantle.

Partial to complete destruction of ferromagnesian silicates, and solution and comminution of the remaining minerals, especially plagio-

clase, have produced a granular sandy aggregate. In the size range between 0.5 mm and 1.0 mm more than half the grains (mostly plagioclase) are subangular to subrounded and show

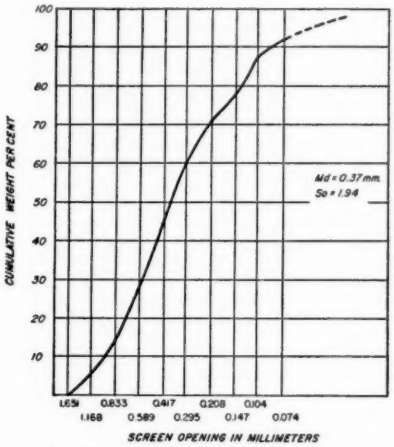


FIGURE 9.—CUMULATIVE DIAGRAM

Screen analysis of loose surface material covering Iron Dike near Sugarloaf. Md = median value. So = sorting coefficient

the effects of ground-water solution of the corners and edges of originally sharply angular, tabular crystals. Samples collected from the surface of the weathered exposure yielded the cumulative size-distribution curve shown in Figure 9, a curve similar to that obtained from many fairly well-sorted, clastic sedimentary accumulations. The screen analysis was made after removing all rock fragments.

SUMMARY OF CONCLUSIONS

(1) The Iron Dike is petrographically unlike and probably older than most of or all the igneous rocks that accompanied and followed Laramide orogeny in the northern Front Range. The dike formed by a single intrusion of an iron-rich diabase magma into a laterally and vertically extensive crustal fissure.

(2) Solidification progressed inward from the chilled, glassy walls of the dike by essentially simultaneous crystallization of pyroxene and labradorite. Magnetite formed before, during, and after crystallization of the pyroxene and labradorite.

(3) Movement in the nearly solidified dike fractured crystals and produced channels which localized chloritization of pyroxene and sericitization of plagioclase by deuteric solutions. Quartz and potassium feldspar are late interstitial minerals.

(4) A soil developed over the dike is a lithosol of the Underwood-Babb type.

(5) Weathering has concentrated limonite at the base of the loose rock cover and in joints in the dike. An increase in ferrous oxide from the base of the cover toward the surface is attributed to the reducing action of organic substances.

(6) Weathering removed CaO and MgO in approximately the same ratios at various depths. The CaO + MgO minimum coincides with the H_2O maximum in the weathered rock. SiO_2 and Al_2O_3 form residual concentrations in the upper part of the mantle.

(7) The mineral most susceptible to weathering in the dike is pyroxene. Sericite in the labradorite has exerted a protective influence preventing alteration. Deuteric minerals, except uralite, also have resisted alteration.

(8) Crystalline clay minerals are practically absent in the weathered rock.

(9) Completely disaggregated rock in the mantle has a grain-size distribution similar to that in some well-sorted sediments.

(10) Deeper weathering in granodiorite next to the dike is attributed to greater permeability.

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UPPER CROIXAN STRATIGRAPHY, UPPER MISSISSIPPI VALLEY

BY C. A. NELSON

ABSTRACT

In the Upper Mississippi Valley Croixan strata above the Franconia formation consist of dolomite and dolomitic siltstone and sandstone overlain by massive nondolomitic sandstone. Previous classifications of this succession have employed both faunal and lithologic criteria for discrimination of formations and members.

The strata can readily be divided on lithologic characters alone, and such divisions are markedly time-transgressive from shoreward to basinward facies. The terminology adopted represents a return to early lithologic classifications in which the St. Lawrence and Jordan formations are recognized as the dolomitic strata and the massive sandstone, respectively. The St. Lawrence is further divided into the Black Earth member (sandy and silty dolomite) and the Lodi member (dolomitic siltstone and fine-grained silty and dolomitic sandstone). Faunal zones are largely independent of the lithologic units.

It is urged that the term Trempealeau, previously employed as a formation name to include all these strata and also as a stage name, be restricted to Trempealeauan Stage.

CONTENTS

TEXT

	Page
Introduction	165
Upper Croixan problem	165
Acknowledgments	167
History of the nomenclature	167
Proposed nomenclature	170
Paleontology	171
Stratigraphy	172
General	172
St. Lawrence formation	173
Definition	173
Black Earth member	173
Lodi member	173
Jordan formation	176
Depositional history	177
Measured sections	177
Osceola	177
Soo Drawbridge	178
Stillwater	178
Afton	179
Red Wing	179
Ridgeland	180
Menomonie Township	180
Oak Hill	181
Spring Green	182
Black Earth	182
References cited	183

ILLUSTRATIONS

Figure		Page
1.	Index map	166
2.	Nomenclature of Upper Croixan strata of Upper Mississippi Valley	168, 169
3.	Relation of Black Earth dolomite, Lodi siltstone, and Jordon sandstone units to faunal subdivisions of the Trempealeauan Stage	170
4.	St. Lawrence to Trempealeau cross section	174
5.	Victory and Wisconsin River Valley cross section	174
6.	St. Croix Valley to Red Wing cross section	175
7.	Dunn County to Trempealeau cross section	175
8.	Monroe and Vernon counties cross section	176
9.	Juneau to Richland counties cross section	176
10.	Central Wisconsin River Valley cross section	176

INTRODUCTION

Upper Croixan Problem

Strata with which this paper is concerned have in the past been called St. Lawrence, Mendota, Trempealeau, Black Earth, Nicollet Creek, and Lodi. Detailed investigations of them were made in western Wisconsin and the St. Croix Valley, measured sections were examined in the area from the Minnesota River Valley on the west to Madison, Wisconsin, and subsurface data from the Minnesota Valley region were evaluated (Fig. 1). Clearly there exists a need both for revision in stratigraphic limits of some units and for changes in terminology.

This study is concerned principally with the more or less dolomitic sediments above the glauconitic sandstone of the Franconia formation and below the fine-grained lower Jordan

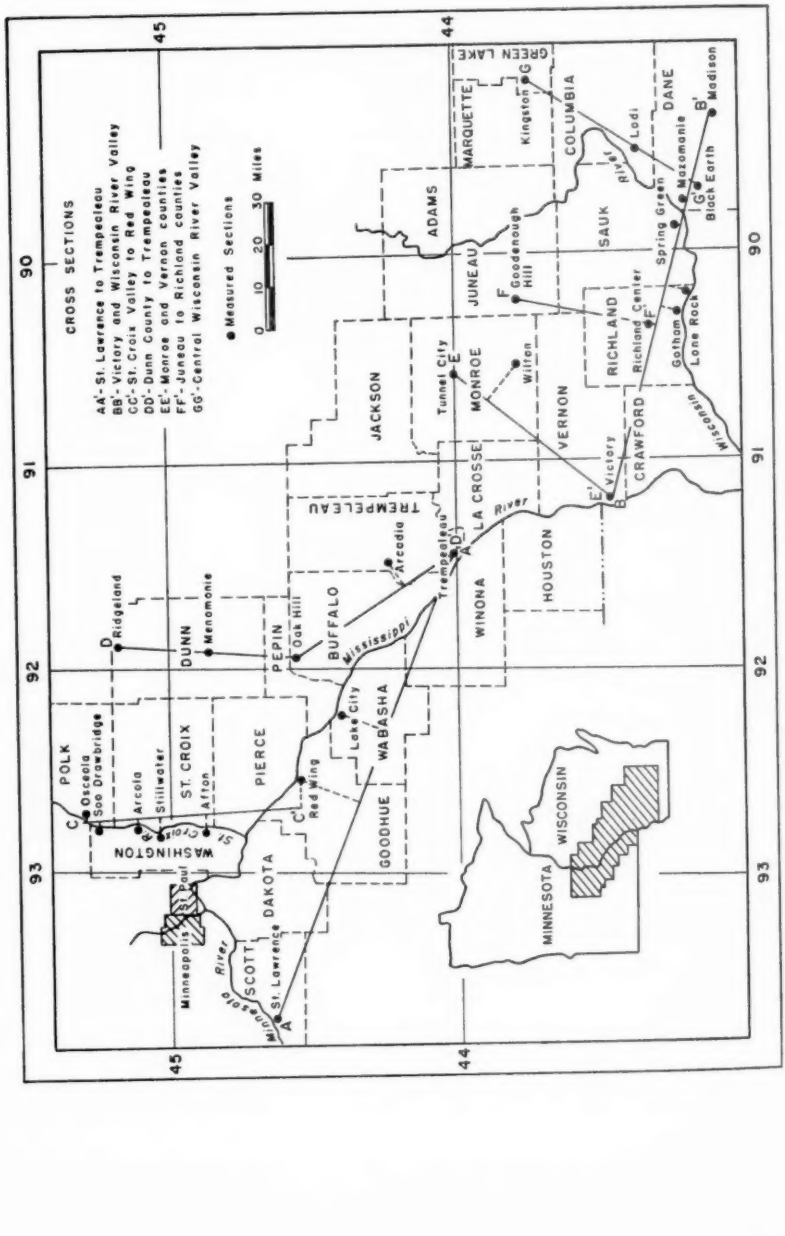

FIGURE 1.—INDEX MAP

sandstone. Although the Jordan unit was not studied in detail over the whole region, suggestions are made concerning its place in the stratigraphic terminology. The larger lithologic units—glauconitic sandstone, dolomitic sediments, and fine-grained sandstone—have been recognized by all investigators since the early studies of Winchell (1874). Commonly, the dolomitic strata have been divided further into two units: a sandy dolomite, called St. Lawrence, Mendota, Black Earth, or Nicollet Creek, overlain by dolomitic siltstone, sandstone, and shale usually called Lodi. Present difficulties stem from the different formational terminologies employed and from minor boundary disputes.

Acknowledgments

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History of the Nomenclature

Figure 2 shows the development of Upper Croixan nomenclature to the present.

In 1874, N. H. Winchell (p. 152–155) described the St. Lawrence limestone, named by Alexander Winchell (1872), as the 14.5 feet of glauconitic and sandy dolomite occurring in quarries at the village of St. Lawrence, Scott County, Minnesota. At the same time (p. 147–152), N. H. Winchell named the Jordan sandstone for exposures at Jordan, Scott County. There is some question whether he (1874, p. 154–155) meant to include within the St. Lawrence the poorly exposed thin shaly beds (Lodi of today) that occupy a stratigraphic position between the limestone of the quarries and the higher Jordan.

Irving (1875, p. 440–443) named the Mendota limestone for exposures at Lake Mendota, Madison, Wisconsin. He included thick-bedded algal dolomite and overlying thinly bedded shaly and arenaceous strata (Lodi). Clearly the Mendota of Irving directly overlies greensands of the present Franconia and underlies the sandstone that he named Madison. Irving equated the Mendota and Madison with the St. Lawrence and Jordan of Winchell.

In 1886, Winchell (p. 334–337) agreed with Irving's correlation and included the "Lodi" strata in the St. Lawrence formation. Although at this time Winchell also considered including in the St. Lawrence the "shaly beds" underlying the quarry rock, he did not include them in his graphic section and formational list, but again drew the lower limit of the formation at the base of the thick-bedded quarry limestone.

Hall and Sardeson (1895, p. 172–174) present conflicting views regarding the base of the St. Lawrence. The thickness they give, 213 feet, suggests inclusion of strata beneath the St. Lawrence quarry rock, but they consider the quarry rock as the lower half of the formation.

The first classification clearly recommending expansion downward to include beds now considered upper Franconia was by Berkey (1897, p. 373). He considered that the St. Lawrence directly overlies the newly named Franconia sandstone and contains dolomite and shale; presumably the shale is the unit now called high Franconia greensand (Berg, 1954). This trend was continued by Winchell in 1905 (p. 268) and by Hall *et al.* (1911, Pl. VI), who went one step further, ignoring Berkey's Franconia, and expanded St. Lawrence downward to the top of the Dresbach sandstone. Hall's Dresbach, and possibly Winchell's, included part of the lower Franconia of today.

Ulrich (*in* Walcott, 1914, p. 354) returned to Berkey's usage and also elevated the top of the St. Lawrence to include strata equivalent to the fine-grained sandstone of the type Jordan. He restricted the Jordan to the coarse-grained sandstone above that of the type section, excluding all the beds that Winchell originally defined as Jordan (Stauffer, 1925, p. 713).

The first return to Winchell's original definition of the base of the St. Lawrence was in 1919 when Twenhofel and Thwaites (p. 616) reported greensand conglomerate below the

dolomite. These they used to identify the base of the St. Lawrence formation and thus expanded the Franconia upward to include the greensand below the conglomerate. They included the fine sandstone of type Jordan in the St. Lawrence, as had Ulrich.

above. Stauffer also (1934, p. 338–342) continued to draw the base of the St. Lawrence at the top of Berkey's Franconia. This was the official position of the Minnesota Geological Survey in 1932 (Grout *et al.*) and was continued by Schwartz in 1936 (p. 31–35).

<table>
<tr><th>WINCHELL (1874, 1886)</th><th>IRVING (1875)</th><th>BERKEY (1897) WINCHELL (1905) HALL et al. (1911)</th><th>ULRICH (Walcott, 1914)</th><th>TWENHOFEL AND THWAITES (1919)</th><th colspan="2">THWAITES (1923) ULRICH (1924)</th><th>STAUFFER (1925, 1934) GROUT (1932) SCHWARTZ (1936)</th><th colspan="2">ULRICH AND RESSER (1930, 1933)</th></tr>
<tr><td rowspan="2">Jordan Sandstone</td><td rowspan="2">Madison Sandstone</td><td rowspan="2">Jordan Sandstone</td><td>Madison fm.(2)</td><td>Madison fm.(2)</td><td colspan="2">Madison fm.(2)</td><td rowspan="2">Jordan Sandstone</td><td colspan="2">Madison fm.(2)</td></tr>
<tr><td>Jordan ss.</td><td>Jordan ss.</td><td colspan="2">Jordan ss.</td><td colspan="2">Jordan ss.</td></tr>
<tr><td rowspan="3">St. Lawrence limestone</td><td rowspan="3">Mendota limestone</td><td rowspan="3">St. Lawrence formation</td><td rowspan="3">St. Lawrence formation</td><td rowspan="3">St. Lawrence formation</td><td rowspan="3">Trempealeau fm.</td><td>Norwalk</td><td rowspan="3">St. Lawrence formation</td><td rowspan="3">Trempealeau fm.</td><td>Norwalk</td></tr>
<tr><td>Lodi</td><td>Lodi</td></tr>
<tr><td>St. Lawrence</td><td>St. Lawrence</td></tr>
<tr><td>Shales</td><td>Potsdam</td><td>(1)</td><td>(1)</td><td>Franconia fm.</td><td colspan="2">Basal(3) Shale
Mazomanie fm.</td><td>(1)</td><td colspan="2">Mazomanie fm.</td></tr>
</table>

FIGURE 2.—NOMENCLATURE OF UPPER CROIXAN STRATA OF UPPER MISSISSIPPI VALLEY
(1) Part of present Franconia formation included in St. Lawrence formation
(2) Madison formation excluded from Cambrian, except by Thwaites (1923)

Thwaites (1923, p. 544–548), announcing a revised nomenclature credited to Ulrich and later published by him (Ulrich, 1924, p. 83), introduced the formational term, Trempealeau, to supplant the St. Lawrence of Twenhofel and Thwaites (1919). Thwaites and Ulrich recognized a Basal Shale, restricted St. Lawrence to the dolomite, and introduced the names Lodi for the dolomitic siltstone and shale, and Norwalk for the fine sandstone of type Jordan. Reasons for a single formational name, Trempealeau, for this diverse rock sequence were the faunal unity (*Saukia* fauna) throughout, gradational relations between highest Lodi and lower Jordan (Norwalk), and the belief, subsequently refuted, that the coarse sandstone which they called Jordan was of continental origin. The Basal Shale of Ulrich included the greensand conglomerate noted earlier and fine-grained shaly sandstone above it.

Stauffer (1925, p. 713) called attention to Ulrich's restriction of Jordan to beds which do not occur at the type locality and argued for a return to a broad interpretation of the Jordan formation to include fine-grained sandstone of the type locality and the coarser sandstone

Ulrich and Resser (1930, p. 11) continued Ulrich's usage of 1924 but did not recognize the Basal Shale member of the Trempealeau formation. Resser (1933) included a basal greensand member below the St. Lawrence dolomite member and also separated the Myers Hill sandstone from the lower part of the Norwalk. This unit was drawn entirely on faunal grounds and did not survive in subsequent classifications.

Martin (1932, p. 4) suggested expansion of Trempealeau to include the coarse-grained upper Jordan, and this practice was continued by Wanenmacher, Twenhofel, and Raasch (1934, p. 21), who recognized the following members: basal greensand and greensand conglomerate, St. Lawrence dolomite, Lodi, and Jordan (including both the Norwalk and coarse-grained sandstone). This represents a transition between the view of Ulrich and the later view of Twenhofel, Raasch, and Thwaites (1935).

After a review of Cambrian problems in the Upper Mississippi Valley, Trowbridge and Atwater (1934, p. 50–65) recommended removal of Norwalk as a member term and a return to the

original usage of Jordan, essentially in accord with Stauffer. For the strata beneath lower Jordan they urged a return to the restricted usage of St. Lawrence, the Winchell position of 1886. Trowbridge and Atwater recognized three members of the St. Lawrence formation: an St. Lawrence and Jordan formations, but used different member subdivisions and retained Madison as a formation.

In an attempt to bring the Minnesota Geological Survey classifications into closer agreement with classification of other Cambrian

MARTIN(1932) (FMS. ONLY) WANENMACHER et al. (1934)		TROWBRIDGE AND ATWATER (1934)		TWENHOFEL RAASCH, AND THWAITES(1935) TROWBRIDGE(1935		BRIDGE (1937)		RAASCH(1939) HOWELL et al. (1944)		STAUFFER et al.(1939) STAUFFER AND THIEL (1941)		RAASCH[4] (1951)		THIS PAPER	
Madison fm.		Jordan formation		formation	Madison	Madison ss.		Madison fm.		Jordan fm.	missing	Sunset Pt. fm.		Sunset Pt. fm.	
Trempealeau formation	Jordan				Jordan	Jordan fm.	Sandstone	Trempealeau formation	Jordan		Van Oser	Trempealeau formation	Jordan	Jordan formation	
							Norwalk				Norwalk		Lodi		
	Lodi	St. Lawrence fm.	Lodi	Trempealeau	Lodi	St. Lawrence fm.	Lodi		Lodi	St. Lawrence fm.	Lodi		St. Lawrence	St. Lawrence fm	Lodi
	St. Lawrence		Black Earth		St. Lawrence		Mendota		St. Lawrence		Nicollet Creek				Black Earth
	Basal[3] congl.		unnamed[3]		Basal[3] congl.		Shale[3]						Arcadia		(5)
Franconia fm.		Franconia fm.		Franconia fm.				Franconia fm.		Franconia fm.		Franconia fm.		Franconia fm.	

(3) Basal beds of Trempealeau or St. Lawrence formation now included in uppermost Franconia formation

(4) Member divisions do not coincide with others because of faunal definition

(5) Inclusion of greensand conglomerate in Franconia formation after Berg (1954)

unnamed member, equivalent to Ulrich's Basal Shale; Black Earth dolomite member, named initially by Ulrich (1916, p. 477–478) and equivalent to Winchell's St. Lawrence limestone (1874); the Lodi member, the dolomitic siltstone and shaly sandstone horizon. This represented the first usage, since Winchell (1886), of a simple two-fold formational nomenclature for the strata above the Franconia greensand: St. Lawrence and Jordan formations.

A further expansion of Trempealeau to include highest Cambrian (Madison of eastern Wisconsin) was made by Twenhofel, Raasch, and Thwaites (1935, p. 1690) with the recognition of five members: basal conglomerate and greensand, St. Lawrence (restricted to the dolomite facies), Lodi, Jordan, and Madison. The same view was presented by the Conference Classification (Trowbridge, 1935, Fig. 1), except that in the accompanying text (Raasch, 1935, p. 310–313) Madison is excluded from the Trempealeau, a practice continued by Raasch in 1939 (p. 115).

Bridge (1937, Table 2) indicated the U. S. Geological Survey classification, which followed Trowbridge and Atwater in recognizing the workers, Stauffer, Schwartz, and Thiel (1939, p. 1228) agreed to draw the base of the St. Lawrence formation above the Franconia greensand and to recognize formally the Lodi member as the strata above the dolomite (original St. Lawrence limestone) which they termed the Nicollet Creek member. Jordan was retained in the original Minnesota usage (all sands between St. Lawrence and the Ordovician dolomite) and was divided into two members: Norwalk and overlying Van Oser.

The Cambrian subcommittee (Howell *et al.*, 1944) repeated the position of Raasch (1939) and recognized St. Lawrence, Lodi, and Jordan as members of the Trempealeau formation, overlain by the Madison formation.

The most recent published classification is in some respects the most confusing. Raasch (1951, p. 147–150) retained his earlier definition of Trempealeau but abandoned the long-accepted lithic subdivisions. He substituted a faunal definition for members and recognized in ascending order: Arcadia, St. Lawrence, Lodi, and Jordan. These names, except for Arcadia, have been used since the work of N. H. Winchell and Ulrich but do not in Raasch's classification

mean what they have meant in the lithologic classifications. His Arcadia member, for example, includes the widely recognized Lodi and Jordan lithologic units (Nelson, 1953, p. 734). Raasch also suggested substitution of the term Sunset Point for Madison.

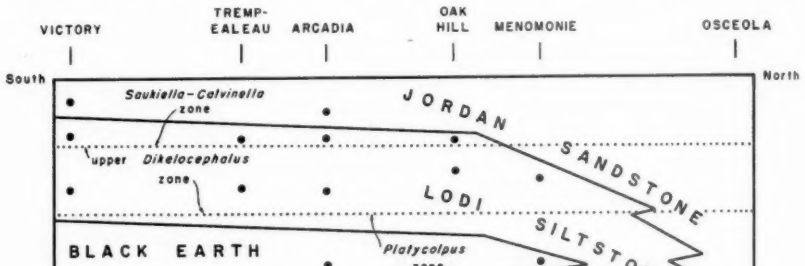

FIGURE 3.—RELATION OF BLACK EARTH DOLOMITE, LODI SILTSTONE, AND JORDAN SANDSTONE UNITS TO FAUNAL SUBDIVISIONS OF THE TREMPEALEAUAN STAGE

Proposed Nomenclature

The post-Franconia sediments have been included either in a two-fold (St. Lawrence and Jordan) or single (Trempealeau) formational classification.

As originally defined by Winchell, the St. Lawrence and Jordan formations were discriminated because of the recognition of two widespread lithologic units—a succession of dolomitic sediments overlain by sandstone.

The present study shows that these lithologic units, though locally variable, are persistent in character and remarkably distinct throughout the region. They have long been used by geologists, drillers, and quarry men. Neither the Norwalk of Ulrich nor the St. Lawrence division has a consistent faunal character. The units climb the biostratigraphic column from northwest to southeast (Fig. 3), and the Lodi-Norwalk boundary lies biostratigraphically lower at Osceola than at Victory. The faunal content of any part or of the whole unit will not lead to a consistent stratigraphic classification; only by reference to the lithologic characters can a reasonable division of the post-Franconia strata be made. Ulrich's Norwalk everywhere represents the initial fine-grained phase of the Jordan clastic depositional cycle and everywhere rests upon carbonate, siltstone, and fine-grained silty sandstone strata. This calcareous and silty unit everywhere lies above terminal beds of the Franconia formation, in many places represented by greensand conglomerate. Thus, by its stratigraphic position between two sandstone units and by its lithologic unity, it is readily recognized and should possess formational status—*St. Lawrence*. In this paper, therefore, the two-fold classification is adopted.

Trempealeau should be abandoned as a formational term and its usage restricted to Trempealeauan Stage. Trempealeau was proposed (Thwaites, 1923; Ulrich, 1924) to supplant, except for the basal portion, the St. Lawrence as used by Ulrich (*in* Walcott, 1914). Ulrich's usage represented an upward expansion of the St. Lawrence of Winchell to include beds equivalent to the type Jordan. Adoption by Ulrich of a single formational term, St. Lawrence in 1914 and Trempealeau in 1924, was based on presumed faunal unity throughout, now well established (but at no time a valid formation criterion), and a belief that the overlying sandstone (Ulrich's Jordan, restricted) was of continental origin. The latter assumption has since been disproved by several discoveries of faunules high in the Jordan.

The Trempealeau, therefore, in its original usage constituted an unnatural grouping of lithologic units and separated the type Jordan from upper Jordan sandstone. This latter ob-

jection was removed by Martin (1932) and by subsequent definitions of Trempealeau expanded to include all the Jordan sandstone. Present usage of Trempealeau, however, relegates St. Lawrence and Jordan to member status and to equal rank with Lodi, a unit clearly included within St. Lawrence and its equivalent (Mendota) by Winchell and Irving. No purpose is served by the retention of Trempealeau. Return to original usage is supported by considerations of priority and by the more accurate discrimination of stratigraphic units which results.

The St. Lawrence formation is divided by the author into two members: The Black Earth and the Lodi. This is the view of Trowbridge and Atwater (1934) except that their basal unnamed member is considered uppermost Franconia (Berg, 1954).

The term Black Earth was first used by Ulrich (1916, p. 477) for dolomite exposed near Black Earth, Wisconsin. It was adopted as a member name by Trowbridge and Atwater (1934, p. 61) because it was the only name that had been applied to the dolomite alone. That adoption is followed here.

In 1939 Nicollet Creek was proposed as a member term by Stauffer *et al.* for the basal St. Lawrence dolomite. Field examination by the author of the type section at Nicollet Creek, Nicollet County, Minnesota (Stauffer and Thiel, 1941, p. 45), discloses that all but the upper 10.5 feet is uppermost Franconia formation. Nicollet Creek can, therefore, be abandoned on grounds of stratigraphy as well as of priority.

The term Lodi was introduced by Ulrich (Thwaites, 1923; Ulrich, 1924, p. 86) for strata between St. Lawrence dolomite and basal Jordan (Norwalk) sandstone, but no specific locality was mentioned as typical. Outcrops near Lodi, Wisconsin, are poor, and the only published section from that locality (Raasch, 1939, Fig. 9) does not include thickness data. Nevertheless, the term Lodi has become so widely established that no useful purpose would be served by the adoption of a new name. Lodi is, therefore, retained for the dolomitic siltstone and sandstone member of the St. Lawrence formation.

The Jordan was studied in detail by the author only in the St. Croix Valley where no consistent member divisions could be discerned. The Norwalk and Van Oser members, defined by Stauffer *et al.*, (1939) are not particularly useful except to emphasize that the Jordan grades upward from fine-grained to coarse-grained. This view is supported by the lack of differentiation of Norwalk and Van Oser in all except two of the measured sections of Stauffer and Thiel (1941). The published sections of Twenhofel, Raasch, and Thwaites (1935) likewise show no consistent vertical divisions within the Jordan unit.

The relation of the Jordan sandstone to the "Madison" sandstone of eastern Wisconsin was not investigated. This relation has been the subject of considerable controversy. If, as some investigators believe, the "Madison" is distinct from the Jordan, it deserves formational status. Raasch (1951, p. 150) has proposed Sunset Point formation, and the author can agree with this proposal. Although the name Madison was proposed for the Cambrian sandstone before its proposal for the Mississippian limestone of the Rocky Mountain region, the latter usage has become well-established in geologic literature. Since the "Madison" sandstone is of limited areal distribution, less confusion will result if it, rather than the Madison limestone, is renamed.

Paleontology

The faunal succession in the St. Lawrence and Jordan formations has been outlined in the Cambrian Correlation Chart (Howell *et al.*, 1944). The zonation of the chart is the same as the proposed *Saukia* zone of Raasch (1951, p. 151) except that the base of the *Saukia* zone lies within the upper beds of the Franconia formation.

The *Saukia* fauna has been partially described by Walcott (1914), Ulrich and Resser (1930; 1933), and Raasch (1939; 1951). Significant revisions in terminology of described species and varieties were made by Raasch (1951).

The faunal zonation used in this paper is as follows, in descending order:

Saukiella-Calvinella zone
  Saukiella subzone
  Calvinella subzone

Upper *Dikelocephalus* zone
Platycolpus zone
Osceolia fauna

The *Plethopeltis* zone (above the *Saukiella-Calvinella* zone) of the Cambrian Correlation Chart was not encountered in the present investigation, since it is essentially restricted to the Sunset Point unit.

The *Osceolia* fauna, defining the Arcadia member of Raasch (1951), is moderately widespread; it has been recognized from the St. Croix Valley to Juneau County. It occurs in both Lodi and Jordan in the northern St. Croix Valley and at Ridgeland, in Lodi only at Menomonie, and in the Jordan in northeastern Buffalo County at Mondovi and Eleva (Ulrich and Resser, 1930, Pls. 17, 19, 23), and at Goodenough Hill. Raasch (1951, p. 145) interprets the latter occurrence as a sandstone facies of the Upper *Dikelocephalus* zone (his Lodi member), although the species represented (*Saukiella pyrene*) is elsewhere characteristic of the *Osceolia* fauna. Berg (1954, p. 878) reports *Osceolia* from basal St. Lawrence (Lodi) at Franklin in southwestern Jackson County, Wisconsin. The species most characteristic of the *Osceolia* fauna are *Osceolia osceola*, *Dikelocephalus thwaitesi*, *Saukiella pyrene*, and *Tellerina leucosia*.

The *Platycolpus* zone (*Platycolpus-Scaevogyra* zone of the Cambrian Correlation Chart), defining the St. Lawrence member of Raasch (1951), is rarely seen to advantage in the region of this study. Fragmentary specimens representative of the zone occur in the Black Earth dolomite at several localities, but it has been definitely recognized only in part of the St. Croix Valley and at Menomonie. The *Platycolpus* fauna occurs generally in dolomite but is found as well in Lodi siltstone at Afton and Arcola. It is characterized by *Platycolpus eatoni* (rarely well preserved) and undescribed species of *Eurekia*, *Dikelocephalus*, and *Finkelnburgia*.

The Upper *Dikelocephalus* zone (Raasch's (1951) Lodi member which he divided into three zonal units) is the most widespread division of the Trempealeauan Stage. The fauna occurs throughout the area generally in Lodi siltstone and shale, although it is found in the Jordan at Tunnel City, Wilton, and Lodi, and has been reported from dolomite at St. Lawrence by Sardeson (1924, p. 119). It is characterized by *Dikelocephalus oweni*, *D. minnesotensis*, *Tellerina crassimarginata*, *Saukia lodensis*, and *Saukiella indenta*.

The *Calvinella* subzone (Raasch's (1951) lower Jordan member) occurs in both the Lodi and Jordan lithologic units. It is characterized by *Calvinella spiniger*, *C. wisconsinensis*, *Dikelocephalus marginatus*, and *Eurekia eos*.

The *Saukiella* subzone (Raasch's (1951) upper Jordan) occurs generally in the Jordan sandstone but is known from the Lodi member at Red Wing and in the region of Lake City and Wabasha, Minnesota. Its fauna is not abundant in numbers of species, but some forms occur profusely. Most common are *Saukiella pepinensis* and *Dikelocephalus marginatus*.

A significant feature of the zonal subdivisions in the St. Lawrence and Jordan formations is the disregard of facies control. Most occur in the dolomitic siltstone facies of the Lodi and in the fine-grained sandstone facies of the Jordan. Exceptions can be seen in Figure 3 and in Figures 5, 7, 8, and 10 which show the general restriction of the *Platycolpus* fauna to the Black Earth member and the *Osceolia* fauna to the siltstone of the Lodi or to higher strata. This suggests that the *Platycolpus* and *Osceolia* zones are in part laterally equivalent biofacies. Few examples (Fig. 3) could illustrate better the impossibility of using faunal criteria to designate members, as Raasch (1951) has suggested.

Stratigraphy

General

In the following stratigraphic discussion, an attempt is made to draw regional conclusions from the detailed study undertaken and from other available data. The author makes use of published measured sections, faunal studies, and other stratigraphic information, credited principally in the cross sections of Figures 4–10. In some instances, published sections have been modified. For example, the basal Jordan, bed 15, of Twenhofel, Raasch, and Thwaites (1935, p. 1737) at Trempealeau which "resembles Lodi but contains *Eurekia*

eos" is here considered part of the Lodi member of the St. Lawrence formation. The same is true of beds 25 and 26 at Victory (p. 1734). Also regarded as lodi are their St. Lawrence member, bed 23 at Tunnel City (p. 1732) and bed 12 at Wilton (p. 1731), which contain no dolomite but are fine-grained dolomitic sandstone and siltstone. Similarly, strata which they include in the Lodi, as bed 8 at Kingston (p. 1723) and beds 15 and 16 at Wilton (p. 1731), are considered basal Jordan formation.

The density of faunal data on the cross sections and in the measured sections indicates that the boundaries of the faunal zones are in most instances arbitrary and in some places necessarily interpretive. Where exposures are good and published information clear, there has been no difficulty in drawing the lithologic boundaries.

St. Lawrence Formation

Definition.—The St. Lawrence formation of the present study consists of the essentially dolomitic sediments between the sandstones of the Franconia and Jordan formations. Its division into Black Earth and Lodi members is based on lithologic characters. It ranges in thickness from 63 feet in a typical basinward exposure at Lake City to 15 feet in the shoreward areas at Ridgeland and Kingston.

Black Earth member.—The basal member of the St. Lawrence is represented by sandy dolomite and interbedded dolomitic siltstone and fine-grained sandstone. In the vicinity of Black Earth and Madison and at localities along the Mississippi Valley it generally is massive, brown to buff, slightly glauconitic, and has algal structures locally. A more distinctly glauconitic facies occurs along the Minnesota River Valley, especially at St. Lawrence where the beds described as the St. Lawrence limestone by Winchell are highly glauconitic and resemble in some respects the upper strata of the Franconia formation (Reno member of Berg, 1954). They are, however, much more dolomitic than typical Reno sediments and lie above essentially nondolomitic Reno greensands capped by characteristic greensand conglomerate.

In its more basinward occurrences, where it rests directly on Franconia greensand (Figs. 4, 5), the Black Earth member contains only small amounts of interbedded dolomitic siltstone and sandstone. Elsewhere, in a shoreward direction (Fig. 7), the Black Earth member has climbed the stratigraphic column and rests on strata of the Lodi member. At these localities considerably greater amounts of siltstone and sandstone occur. The occurrence of basal Lodi strata beneath the Black Earth dolomite (Figs. 7, 8) indicates the initiation of moderate carbonate development before the transgression that brought the basinward facies to the north.

The Black Earth member contains algal structures and a few representatives of the *Platycolpus* zone. Where the evidence is available the member appears to coincide with the *Platycolpus* zone. The dolomite beds of the Black Earth member probably approximate a time horizon and mark the maximum transgression of the Croixan sea.

The sand-silt content of the basal St. Lawrence is shown approximately by percentages of insolubles. At basinward localities Stauffer and Thiel (1933, p. 65–74) show a range from 7.1 to 9.08 per cent insolubles in three samples from the Minnesota Valley region; Irving (1877, p. 543) reports 7.8 per cent in a sample from Madison. Analyses made during the present study show a sand-silt content of 12.4 per cent at Black Earth, rather higher than for typical basin facies dolomite. The higher sand-silt content of the shoreward facies (Fig. 7) is shown by percentages of 28.3 and 38.1 at Menomonie and Oak Hill, respectively.

The Black Earth member ranges in thickness from 8 to 13 feet in the basinward facies, as at St. Lawrence, Lake City, Gotham, Spring Green, and Madison. Shoreward to the north the member thins to 0.5–3 feet at Goodenough Hill, Wilton, and Menomonie. The presence of 2.5 feet, high in sand-silt content, at Red Wing reflects the high clastic and low carbonate nature of the St. Lawrence formation in the St. Croix Valley and vicinity.

Lodi member.—The Lodi member consists of siltstone, generally dolomitic, and dolomitic sandstone above the Franconia and beneath the Jordan. Lodi strata vertically succeed Black Earth dolomite beds in the basinward exposures (Figs. 4, 5). Elsewhere the Lodi

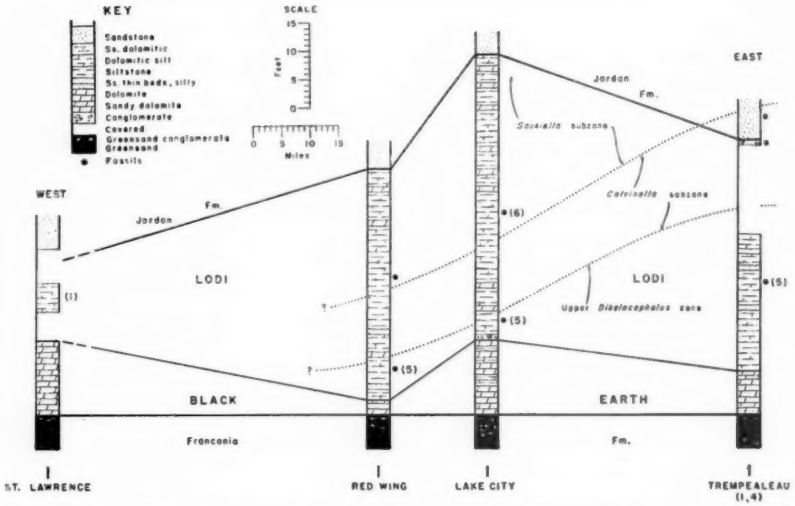

FIGURE 4.—ST. LAWRENCE TO TREMPEALEAU CROSS SECTION

Lake City section from W. C. Bell (personal communication)
Nos. (1) to (6) on Figures 4–10 indicate data from following sources:
(1) Twenhofel, Raasch, and Thwaites (1935)
(2) Raasch (1939)
(3) Raasch (1951)
(4) Ulrich and Resser (1933)
(5) Ulrich and Resser (1930)
(6) Stauffer and Thiel (1946)

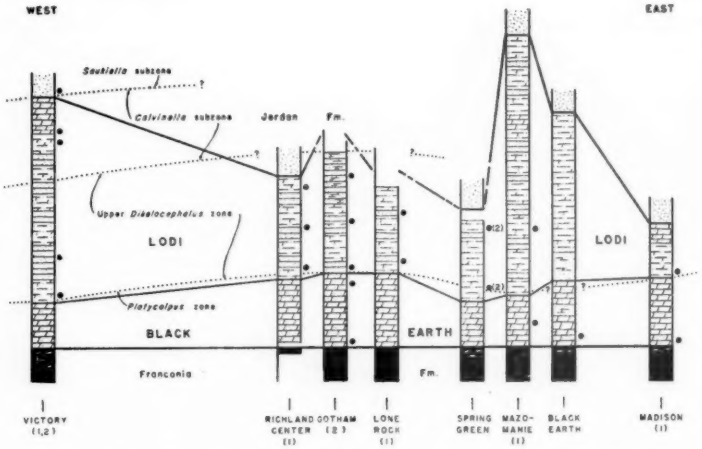

FIGURE 5.—VICTORY AND WISCONSIN RIVER VALLEY CROSS SECTION

See Figure 4 for references, scale, and lithologic key

occurs beneath the Black Earth member as well as above it (Figs. 7, 8). At Arcadia, these strata beneath the dolomite are regarded by Raasch as type Arcadia member (1951, p. 147). Field examination of these beds reveals that, except for their content of the *Osceolia* fauna, on which basis they were designated Arcadia member, they do not differ from the strata

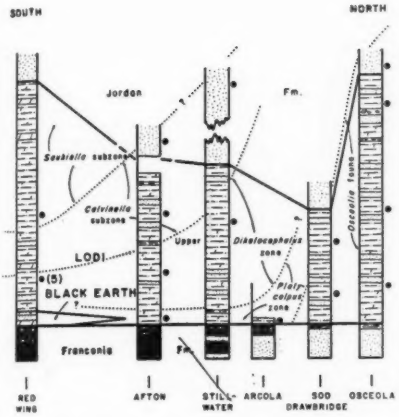

FIGURE 6.—ST. CROIX VALLEY TO RED WING CROSS SECTION
See Figure 4 for references, scale, and lithologic key

Like the Black Earth member, the Lodi exhibits facies change from the basinward exposures to those in shoreward areas to the north. The lithologic change can be seen in the presence of dolomite beds and the higher content of dolomitic siltstone at localities such as Madison and Victory in contrast to localities northward. At Victory, the Lodi is terminated by 6 feet of sandy dolomite included in the basal Jordan by Twenhofel, Raasch, and Thwaites (1935, p. 1734). The beds belong lithologically with the Lodi and are so grouped in Figures 5 and 8. The presence of a basal *Calvinella* subzone fauna (Raasch, 1939, Fig. 7) including *Eurekia eos* probably led these authors to bend the lithic definitions to fit the faunal content.

In the south-to-north-trending sections (Figs. 6–8, 10) the percentage of clastics in the Lodi

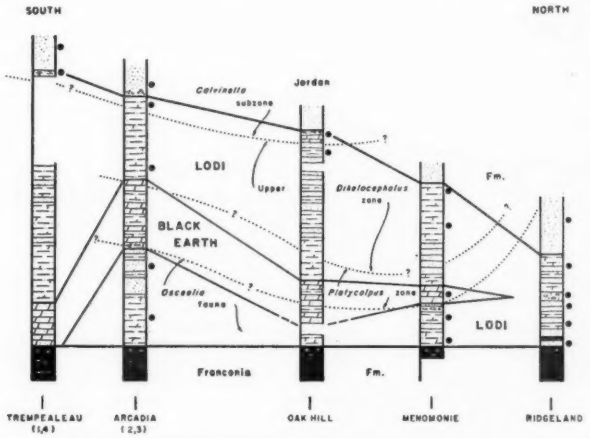

FIGURE 7.—DUNN COUNTY TO TREMPEALEAU CROSS SECTION
See Figure 4 for references, scale, and lithologic key

above the dolomite. These upper beds Raasch calls Lodi (1939, Fig. 8), presumably because they contain the Upper *Dikelocephalus* zone fauna. North from Arcadia, the basal Lodi strata can be recognized beneath Black Earth dolomite at Oak Hill and Menomonie (Fig. 7) until the dolomite disappears, as at Ridgeland, where the Lodi is continuous from the Franconia to the Jordan. Similar relations occur in the section from Victory to Tunnel City (Fig. 8) and in the St. Croix Valley region (Fig. 6).

increases shoreward. This is apparent in the greater number of sandstone interbeds and is shown in the St. Croix Valley region by the decrease of carbonate in the siltstone at Stillwater and Osceola compared with the sections at Red Wing and Afton.

Analyses record a similar facies change in the Lodi. Stauffer and Thiel (1933, p. 65–74) report an insoluble percentage of 26.4 for the basinal facies near St. Lawrence, contrasting with 52.2 per cent shoreward at Red Wing.

Analyses made during the present study reveal sand-silt percentages from 51.6 to 75.8 at Afton, 64.2 to 89.9 at Stillwater, and from 78.0 to 84.4 at Osceola. This increase in the sandy facies northward in the St. Croix Valley and also in the Dunn County section (Fig. 7) is related to early true Jordan deposition in these areas. Lying close to exposures of Precambrian rocks at present and presumably also during the Croixan, these areas were sites of shore facies sand accumulation throughout much of St. Lawrence deposition.

The Lodi member is characterized by faunas of the Upper *Dikelocephalus* zone, although it contains the *osceolia* fauna at some localities. The most prolific horizons occur in thinly bedded siltstone. Locally, as at Red Wing, Afton, Lake City, and Victory, the Lodi contains the higher faunas of the *Saukiella-Calvinella* zone.

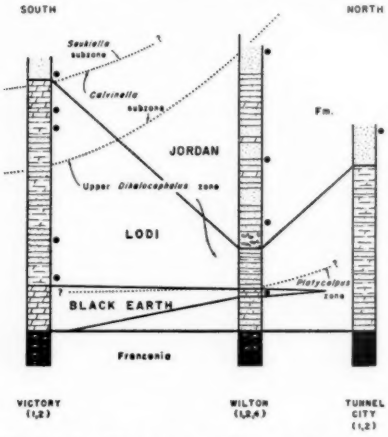

FIGURE 8.—MONROE AND VERNON COUNTIES CROSS SECTION
See Figure 4 for references, scale, and lithologic key

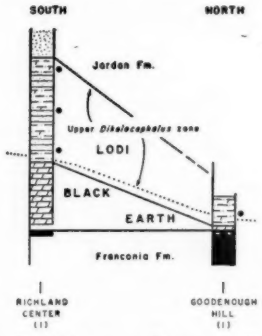

FIGURE 9.—JUNEAU TO RICHLAND COUNTIES CROSS SECTION
See Figure 4 for references, scale, and lithologic key

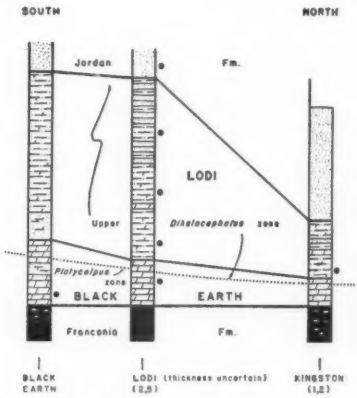

FIGURE 10.—CENTRAL WISCONSIN RIVER VALLEY CROSS SECTION
See Figure 4 for references, scale, and lithologic key

Jordan Formation

The Jordan formation consists of massive to well-bedded, fine- to coarse-grained sandstone that lies above dolomitic strata of the Lodi. The contact between Lodi member and basal Jordan is generally sharp. Though locally thin transitional strata are present, in most localities the boundary can be drawn above the uppermost siltstone and at the base of continuous sandstone.

Figures 3 and 6–8 show the marked biostratigraphic climb of the Jordan from north to south. No independent evidence, such as thin continuous marker beds, traceable bedding planes, or tuffaceous beds is available as a time horizon. Although the biostratigraphic units (zones and subzones) are probably somewhat time-transgressive, they more closely approximate lines of synchrony than do the lithologic units. They are here, therefore, regarded as time markers. This view is supported in part by the lack of facies control of the zonal units.

At Osceola (Figs. 3, 6) and Ridgeland (Fig. 7) the Jordan contains the *Osceolia* fauna. In the St. Croix Valley the basal Jordan contains the *Osceolia* fauna as far south as Marine, Minnesota, approximately 5 miles south of the Soo

Drawbridge section. The Jordan at Mondovi and Eleva, Wisconsin, 15 and 25 miles, respectively, east of Oak Hill in northeastern Buffalo County, is also reported by Ulrich and Resser (1930, Pl. 17, 19, 23; 1933, Pl. 36) to contain the *Osceolia* fauna. On the other hand, nearby at Durand, 3 miles northwest of Oak Hill (Ulrich and Resser, 1933, Pl. 37), and at Oak Hill, the Jordan lies above the Upper *Dikelocephalus* fauna. Similar relations exists in Monroe and Juneau counties. At Tunnel City and Wilton, the Jordan contains the Upper *Dikelocephalus* fauna, but at Goodenough Hill it contains *Saukiella pyrene*, a species common in the *Osceolia* fauna. Raasch (1951, p. 145) interprets the latter occurrence, reported from Jordan (Norwalk) sandstone by Ulrich and Resser (1933, Pl. 35), as a sandstone facies of the Upper *Dikelocephalus* zone (his Lodi member).

Osceola and northern Dunn County, northeastern Buffalo County, and southwestern Juneau County lie near exposures of Precambrian rocks and represent local southward-projecting tongues of the near-shore facies. At these localities, sand deposition (Jordan formation) began early in the late Croixan and continued until replaced by Oneota (Ordovician) dolomite deposition.

The arrival of the sand facies in more basinal areas was postponed until Late *Dikelocephalus* "time" (Tunnel City, Wilton, and Lodi) and in most places until *Calvinella* "time" (Stillwater, Afton, Arcadia, and Trempealeau). Locally, the beginning of sand facies deposition was even somewhat later. At Red Wing and Lake City, sand deposition began after development of the *Saukiella* subzone fauna, for there *Saukiella* occurs within the Lodi. At Victory, the *Saukiella* fauna occurs at the base of the Jordan formation.

Depositional History

The post-Franconia Cambrian sediments of the region represent a single transgressive-regressive cycle succeeding Franconia sand deposition and preceding deposition of Ordovician dolomite. At the time of the most shoreward advance of the Black Earth facies, as at Menomonie (Fig. 7) and Wilton (Fig. 8), a sandy facies of the Lodi was developing near by at Ridgeland and Tunnel City. Presumably, in the more shoreward areas now lost by erosion, the St. Lawrence facies thinned to disappearance, and Franconia sands were overlain directly by Jordan sandstone. This is suggested by the shoreward thinning of the St. Lawrence interval and its replacement by the Jordan sands (Figs. 7–8, 10). Basinward, the Franconia, St. Lawrence, and Jordan must become increasingly dolomitic. This trend is shown in the subsurface sections (Twenhofel, Raasch, and Thwaites, 1935, p. 1718–1720) from southeastern Wisconsin, where the Lodi and Jordan are represented by considerable dolomite. In the sections, the Black Earth dolomite member thickens at the expense of the Franconia which is moderately dolomitic and reduced to 50–100 feet in thickness. In the outcrop area to the northwest, the Franconia is only locally dolomitic and ranges in thickness from 150 to 210 feet (Berg, 1954, p. 860).

Subsurface information (well cuttings and electric logs) from the region 25 to 30 miles south of St. Lawrence in Le Sueur County, Minnesota, also reveals an increase in carbonate content and a thickening of the St. Lawrence interval basinward to 80–95 feet. Examination of well cuttings reveals that the Lodi interval becomes increasingly dolomitic basinward and probably cannot be separated from the Black Earth dolomite.

Measured Sections

The following sections are representative of the St. Lawrence formation in the region studied. Most are previously undescribed. Sections previously described have been redescribed and reinterpreted in light of the lithologic facies discussed above. They are arranged in order from northwest to southeast.

Osceola

Section at Cascade in Osceola, Polk County, Wisconsin

	Feet
Jordan Formation	
10. Sandstone, buff and brown, fine-grained, massive. Fauna mainly from equivalent strata near by—*Osceolia osceola, Tellerina leucosia, Saukiella pyrene, Dikelocephalus thwaitesi, Illaenurus quadratus*	4.0

St. Lawrence Formation—43.5 feet

	Feet
Lodi member—43.5 feet	
9. Siltstone, sandy, gray to buff, thin-bedded, with interbeds of brown to buff fine-grained dolomitic sandstone. *Osceolia osceola, Tellerina leucosia, Dikelocephalus* sp.	6.0
8. Conglomerate, buff flat sandy siltstone pebbles in brown dolomitic sandstone matrix	1.0
7. Sandstone, buff to brown, fine-grained, thinly laminated, dolomitic toward top	1.5
6. Siltstone, gray and buff, dolomitic, thin-bedded	2.5
5. Conglomerate, flat gray to buff fine-grained sandstone pebbles in dark brown dolomitic sandstone matrix	0.5
4. Siltstone, dolomitic at top, gray and buff, some mottled, and interbedded buff fine-grained slightly glauconitic sandstone, thin- to medium-bedded	18.0
3. Siltstone, dolomitic, light gray, very thin-bedded, with interbedded fine-grained glauconitic sandstone. *Osceolia osceola*	14.0

Franconia Formation

Mazomanie member	
2. Conglomerate, brown sandstone pebbles in slightly glauconitic sandstone matrix and interbedded glauconitic dolomite and brown dolomitic sandstone	1.0
1. Sandstone, somewhat dolomitic, buff to brown, fine- to medium-grained, massive to medium-bedded and gray and buff sandy shale	27.0

Soo Drawbridge

Section at Falls West of Water Tank and Drawbridge, Sec. 7, T. 32 N., R. 19 W., Washington County, Minnesota

Jordan Formation

9. Sandstone, white to brown, fine-grained, thin-bedded (not measured)

St. Lawrence Formation—20 feet

Lodi member—20 feet	
8. Siltstone, buff, sandy, with interbeds of silty and shaly fine-grained sandstone	1.5
7. Sandstone, buff to brown, fine-grained, shaly, thinly laminated	1.0
6. Siltstone, buff to gray, thin-bedded, with interbeds of buff to brown fine-grained sandstone and gray-green shale. *Osceolia osceola, Dikelocephalus thwaitesi*	9.5
5. Siltstone, buff, thin-bedded, and interbedded fine-grained silty sandstone and green shale. *Osceolia osceola*	8.0

Franconia Formation

Mazomanie member	
4. Conglomerate, flat siltstone pebbles in glauconitic and dolomitic matrix	0.5
3. Sandstone, brown, medium-grained, dolomitic	1.0
2. Sandstone, white, medium- to coarse-grained, massive	4.0
1. Sandstone, white to buff and brown, medium- to coarse-grained	7.0

Stillwater

Section in Fairy Glen, NE 1/4 Sec. 21, T. 30 N., R. 20 W., North Edge of Stillwater, Washington County, Minnesota

Jordan Formation

12. Sandstone, white, fine-grained, cross-stratified and some interbedded buff siltstone (not measured)

St. Lawrence Formation—28 feet

Lodi member—28 feet	
11. Shale, silty, gray to buff, and interbedded fine-grained gray shaly sandstone. Sharp contact with Jordan sandstone	3.0
10. Siltstone, gray, massive	1.5
9. Shale, gray to buff, sandy and silty	4.0

	Feet
8. Siltstone, sandy, gray, massive with conchoidal fracture. *Dikelocephalus minnesotensis, D. oweni, Tellerina crassimarginata, Illaenurus calvini*	2.0
7. Siltstone, dolomitic and sandy, mottled buff to gray, with interbeds of gray shale	8.0
6. Sandstone, shaly, buff, fine-grained, thin-bedded	1.5
5. Siltstone, sandy, mottled buff to brown, thin-bedded, with shaly and dolomitic sandstone interbeds. 3 inch dolomitic siltstone conglomerate 2 feet above base	5.5
4. Conglomerate, dolomitic siltstone flat pebbles in glauconitic and dolomitic fine-grained sandstone matrix	0.5
3. Siltstone, dolomitic, mottled buff to gray-green, medium-bedded, with interbeds of gray-green shale and fine-grained glauconitic sandstone	2.0

Franconia Formation

Reno and Mazomanie members

2. Conglomerate, siltstone pebbles in dark glauconite matrix	1.0
1. Sandstone, glauconitic, gray to buff, fine-grained, with interbedded fine-grained dolomitic sandstone	17.0

Afton

Section along Creek, SE 1/4 Sec. 22, T. 28 N., R. 20 W., at South Edge of Afton, Washington County, Minnesota

Jordan Formation

14. Sandstone, yellow to white, fine to medium-grained, thin-bedded (not measured). *Calvinella spiniger, Eurekia* sp.

St. Lawrence Formation—26.5 feet exposed

Lodi member—26.5 feet exposed

13. Covered	3.0
12. Siltstone, dolomitic, gray, thin-bedded	1.5
11. Conglomerate, dolomitic, gray to buff dolomitic siltstone flat pebbles in brown-buff dolomitic siltstone matrix	0.5
10. Siltstone, dolomitic, mottled gray, lenticular bedding. *Calvinella wisconsinensis, Dikelocephalus marginatus* (fauna from nearby section)	2.5
9. Siltstone, gray, massive, conchoidal fracture	0.5
8. Siltstone, dolomitic, gray, thin-bedded, with fine-grained gray sandstone interbeds	7.0
7. Conglomerate, dolomitic, buff to gray flat siltstone pebbles in buff dolomitic siltstone matrix	0.5
6. Siltstone, dolomitic, gray, thin-bedded, with fine-grained gray sandstone interbeds. *Aglaspis* sp.	6.5
5. Sandstone, yellow buff, very fine-grained	0.5
4. Siltstone, dolomitic, gray, thin-bedded	4.0
3. Siltstone, dolomitic, gray, massive to medium bedded. *Dikelocephalus* sp., *Eurekia* sp., *Platycolpus* sp., *Aglaspis* sp., *Finkelnburgia* sp., numerous graptolites	3.0

Franconia Formation

Reno member

2. Conglomerate, glauconitic at base becoming dolomitic at top	1.5
1. Greensand, fine-grained, locally burrowed	52.0

Red Wing

Section at La Grange Mountain (Barn Bluff), Red Wing, Goodhue County, Minnesota

Jordan Formation

11. Sandstone, yellow to buff, fine-grained, massive (not measured)

St. Lawrence Formation—43 feet

Lodi member—40.5 feet

10. Siltstone, dolomitic, yellow to buff, sandy dolomite, and buff dolomitic shale	21.5
9. Siltstone, dolomitic, blue gray to buff, yellow weathering, thin-bedded. *Saukiella pepinensis, Dikelocephalus marginatus*	4.0

	Feet
8. Siltstone, dolomitic, gray, and interbedded buff silty dolomite, with fine-grained sandstone interbeds	12.0
7. Sandstone, buff, fine-grained, with gray shale at base	1.0
6. Siltstone, dolomitic, mottled buff to gray	2.0

Black Earth member—2.5 feet

5. Dolomite, silty, brown	0.5
4. Sandstone, buff to brown, fine-grained, with thin laminae of glauconitic sandstone at base	0.5
3. Sandstone, dolomitic and glauconitic, fine-grained, conglomeratic	0.5
2. Dolomite, brown, crystalline, with dolomitic brown siltstone at top	1.0

Franconia Formation

Reno member

1. Greensand, fine-grained, with greensand conglomerate throughout. Upper 3 feet with buff siltstone and fine-grained sandstone pebbles in dolomitic greensand matrix	10.5

Ridgeland

Roadcuts on State Highway 25, NW 1/4 Sec. 20, T. 31 N., R. 12 W., 3 miles south of Ridgeland, Dunn County, Wisconsin

Jordan Formation

9. Sandstone, white to buff, fine-grained, micaceous, thin-bedded. *Saukiella pyrene, Tellerina leucosia, Illaenurus quadratus*	4.5
8. Sandstone, white to brown, fine-grained, thin-bedded, earthy and dolomitic at base	6.0

St. Lawrence Formation—15.5 feet

Lodi member—15.5 feet

7. Siltstone, dolomitic, gray to buff, thin-bedded, with interbeds of buff to brown very fine-grained sandstone. Conglomeratic near base. *Osceolia osceola, Dikelocephalus thwaitesi, Illaenurus* sp.	6.0
6. Conglomerate, flat buff siltstone pebbles in dolomitic and glauconitic fine-grained sandstone matrix. *Osceolia osceola*	1.5
5. Sandstone, buff, very fine-grained, silty and micaceous, thin-bedded, and buff thin-bedded dolomitic siltstone. *Osceolia osceola, Dikelocephalus thwaitesi*	6.5
4. Greensand conglomerate, large flat siltstone pebbles in greensand matrix	0.5
3. Sandstone, yellow brown, fine-grained, micaceous and thin-bedded, very slightly glauconitic. *Osceolia osceola, Illaenurus quadratus*	1.0

Franconia Formation

Reno member

2. Greensand conglomerate, flat siltstone pebbles in dolomitic greensand matrix	1.0
1. Greensand, dark green to red-brown, medium- to coarse-grained	15.0

Mazomanie member (not measured)

Menomonie Township

Section in Quarry, East Side of County Road K, SW 1/4 Sec. 18, T. 28 N., R. 13 W., Dunn County, Wisconsin. Beds 10 and 11 from Road Cuts along County Road K, SE 1/4 NE 1/4 Sec. 7, T. 27 N., R. 13 W., Dunn County, Wisconsin

Jordan Formation

11. Sandstone, buff to brown, fine-grained, thin-bedded	3.0

St. Lawrence Formation—28.5 feet

	Feet
Lodi member—18 feet	
10. Siltstone, dolomitic, buff, rough blocky to conchoidal fracture. *Illaenurus* sp., *Dikelocephalus* sp.	5.0
9. Siltstone, dolomitic, buff, thin-bedded, upper beds somewhat micaceous. *Illaenurus* sp.	5.5
8. Sandstone, white to yellow-brown, very fine-grained, thin-bedded, micaceous	5.5
7. Siltstone, dolomitic, buff to gray, thin and irregularly bedded	2.0
Black Earth member—3 feet	
6. Dolomite, silty, brown, fine-grained, slightly glauconitic, thin- to medium-bedded, with buff siltstone interbeds. *Eurekia* sp.	3.0
Lodi member—7.5 feet	
5. Siltstone, dolomitic, buff	0.5
4. Conglomerate, flat siltstone pebbles in buff dolomitic and glauconitic fine-grained sandstone matrix	1.0
3. Siltstone, dolomitic, buff, thin-bedded, with buff fine-grained dolomitic sandstone interbeds. *Osceolia osceola, Dikelocephalus thwaitesi*	3.5
2. Siltstone, buff-gray, shaly and fissile, dolomitic at base. *Osceolia osceola*	2.5

Franconia Formation

Reno member	
1. Greensand, dolomitic and conglomeratic. Exposed at floor of quarry	2.0

Oak Hill

Quarry Section on County Road FF, NE 1/4 Sec. 3, T. 24 N., R. 13 W., 1/4 Mile Northeast of Oak Hill School, at Intersection of County Roads FF and F, Buffalo County, Wisconsin

Jordan Sandstone

16. Sandstone, yellow-brown, fine-grained, massive, dolomitic at base, case-hardened	4.0

St. Lawrence Formation—37.5 feet

Lodi member—26 feet	
15. Dolomite, sandy, brown, fine-grained, massive. Upper 9 inches a conglomerate with yellow siltstone and fine-grained sandstone pebbles in dolomitic sandstone matrix. *Eurekia* sp., *Dikelocephalus* sp.	2.0
14. Sandstone, brown, fine-grained, with siltstone pebble conglomerate at base	1.0
13. Siltstone, buff, thin-bedded, rough conchoidal fracture, with interbeds of white fine-grained sandstone. *Illaenurus* sp.	1.5
12. Sandstone, buff to white, fine-grained	1.0
11. Covered	1.5
10. Sandstone, brown to buff, fine-grained, thin-bedded and micaceous, with interbeds of gray and buff siltstone	7.0
9. Siltstone, dolomitic, buff to gray-green, thin and irregularly bedded, with few interbeds of buff fine-grained sandstone	3.0
8. Sandstone, buff to brown, fine-grained, well bedded	1.5
7. Siltstone, dolomitic, buff to gray, thin and irregularly bedded, with interbedded sandy siltstone	7.5
Black Earth member—7.5 feet exposed	
6. Dolomite, mottled buff and gray, heavy-bedded, and interbedded buff dolomitic siltstone. Greensand laminae at base	2.0
5. Dolomite, silty, brown, massive, with interbedded buff to brown fine-grained dolomitic sandstone, some beds conglomeratic	5.5
Lodi member—2 feet exposed	
4. Covered	2.0
3. Siltstone, dolomitic, mottled buff and gray, thin-bedded	2.0

Franconia Formation

	Feet
Reno member	
2. Greensand conglomerate, in part dolomitic	1.5
1. Greensand (not measured)	

Spring Green

Section in Road Cut on State Highway 23, 1.6 Miles North of Spring Green, Sauk County, Wisconsin

Jordan Formation

8. Sandstone, buff, fine- to medium-grained, thin- to medium-bedded (not measured)

St. Lawrence Formation—22 feet exposed

Lodi member—14 feet exposed

7. Covered	2.0
6. Siltstone, dolomitic, gray, thin and irregularly bedded	1.5
5. Siltstone, dolomitic, buff, thin and irregularly bedded, and interbeds of buff, fine-grained sandstone	3.0
4. Siltstone, dolomitic, gray, thin and irregularly bedded	4.0
3. Siltstone, dolomitic, buff, thin-bedded	5.5

Black Earth member—8 feet

2. Dolomite, sandy, buff to brown, massive	8.0

Franconia Formation

Reno member

1. Greensand, conglomeratic at top (not measured)

Black Earth

Road Cut on U. S. Highway 14, 1 Mile East of Black Earth, Dane County, Wisconsin

Jordan Formation

13. Sandstone, white, medium-grained, massive and cross-stratified (not measured)

St. Lawrence Formation—41 feet

Lodi member—29.5 feet

12. Siltstone, dolomitic, mottled red to purple, thin-bedded	4.0
11. Siltstone, dolomitic, mottled yellow brown, thin-bedded to massive	10.0
10. Siltstone, dolomitic, alternating red and purple, thin- to medium-bedded, and interbeds of yellow-brown dolomitic siltstone and buff dolomite	5.5
9. Siltstone, very dolomitic, buff, massive	5.0
8. Sandstone, dolomitic, very fine-grained, yellow, massive	2.0
7. Sandstone, dolomitic, fine-grained, yellow, thin-bedded	3.0

Black Earth member—11.5 feet

6. Dolomite, brown, cavernous, massive, glauconitic, with interbeds of massive dolomitic sandstone	4.0
5. Sandstone, dolomitic, fine-grained, mottled yellow brown, conglomeratic at top	2.0
4. Dolomite, sandy, buff, massive, with irregular and discontinuous lenses of glauconitic and dolomitic sandstone at base	4.5
3. Dolomite, buff, massive, algal	1.0

Franconia Formation

	Feet
Reno member	
2. Greensand, dolomitic, lenticular siltstone conglomerate at top	2.5
1. Greensand (not measured)	

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PEGMATITES IN WEST GREENLAND

BY HANS RAMBERG

ABSTRACT

Numerous quartz-feldspar pegmatites occur in the gneisses and crystalline schists of West Greenland. Some of these pegmatites are volume-by-volume replacement bodies, others formed by slow growth in dilating fissures. The nondilation and the dilation pegmatites were emplaced by similar mechanism—*viz.*, diffusion through the host rocks and through fissures. Field evidence is against bodily flow of pegmatite magma and crystallization from a stagnant magma. Tension joints and shear zones have commonly localized pegmatites, but other structural heterogeneities such as pre-pegmatitic dikes and inclusions in gneisses were also "sinks" for the diffusion currents of pegmatite-forming matter.

Ptygmatic folding and pinch-and-swell structures are results of tectonic deformation simultaneous with—or later than—the formation of pegmatite.

CONTENTS

TEXT

	Page
Introduction	185
Acknowledgments	187
Composition of the pegmatites	187
Distribution and localization	188
Size and shape of pegmatites	191
General	191
Sheet-shaped pegmatites	192
Ptygmatically folded pegmatites	192
Pinch-and-swell structure	193
Pegmatites with irregular shape	195
Pegmatite fillings between boudins	195
Pegmatites wrapped around basic inclusions	197
Structural relations between pegmatite and country rock	198
Unaltered host-rock fragments in pegmatites	200
Internal texture and structure of pegmatites	201
Mineralogical and chemical interrelation between pegmatites and host rock	204
Genesis	208
References cited	213

ILLUSTRATIONS

Figure	Page
1. Map of West Greenland	186
2. Pegmatites in amphibolite	189
3. Pegmatites in amphibolite	189
4. Pegmatites in amphibolite	189
5. Pegmatites in ultrabasics	190
6. Pegmatites in amphibolite and mica schist	190
7. Folded quartzite layer in mica schist	193
8. Pegmatite in gneiss	195
9. Pegmatite in gneiss	198
10. Inclusion with pegmatite in gneiss	198
11. Pegmatite in dunite	202
12. Zoned pegmatite	203
13. Pegmatite in gneiss	203
14. Cross sections of pegmatites	204
15. Pegmatites in boundinage structure	206
16. Feldspar pegmatite in amphibolite	207
17. Contact of pegmatite	207

Plate	Following page
1. Pegmatites 2. Pegmatites 3. Pegmatites 4. Pegmatites and "veins"	190
5. Pegmatites 6. Pegmatites 7. Pegmatites 8. Pegmatites	196
9. Pegmatites 10. Pegmatites; tourmaline vein 11. Pegmatites	202

INTRODUCTION

During 5 summer mapping and field studies in West Greenland, beginning in 1946, the writer studied thousands of pegmatite bodies throughout the exceptionally well exposed Precambrian gneiss and crystalline schist terrain along the coast from Jacobshavn Isfjord at about 69°N. to the famous Ivigtut cryolite mine at 61°N. Much of this coastal area has been mapped systematically by the Greenland Geological Survey, Copenhagen, of which the writer has been a co-worker from the start of the mapping project in 1946. A map with general lithologic-structural description and interpretation of the area is now being prepared for publication. The present paper deals exclusively with pegmatites of the area and structural and lithologic relations relevant to the pegmatite problem.

The investigated coastal part of Greenland is a typical gneiss terrain consisting of the

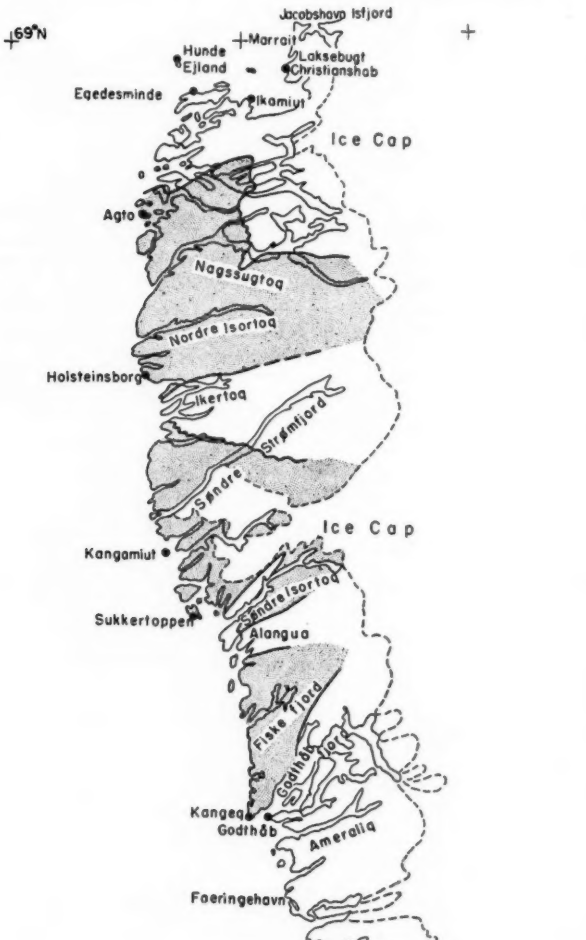


FIGURE 1.—MAP OF WEST GREENLAND

Area throughout which most of the studied pegmatites are distributed. Dotted regions: granulite-facies rocks; blank regions: amphibolite- and in part epidote amphibolite-facies rocks. The northern border of the granulite-facies complex around the outlet of Söndre Strömfjord coincides with the boundary between the old Godthåbidian orogen to the south and the younger Nagssugtogidian orogen to the north.

deeply eroded roots of two Precambrian orogens. The oldest orogen, which occupies the region south of Söndre Strömfjord (Fig. 1), strikes generally north-northeast. It has been called the Godthåbides, which may be identical to the Ketilides of South Greenland (Wegmann, 1938). From Söndre Strömfjord the roots of a younger orogen, the Nagssugtoqides, extend northward beyond Jacobshavn Isfjord. The general strike of the Nagssugtoqides trends east-northeast. As in other deeply eroded folded mountain chains, quartz-feldspar gneisses of various composition and origin are the dominant rocks in the Greenlandic orogens in which metasediments, metavolcanites, amphibolites and ultrabasics form layers, lense

or other remnants in various stages of alteration. Large portions of the oldest orogen, the Godthåbides, are densely penetrated by unaltered steep diabase dikes. In the younger Nagssugtoqidian region the same diabase dikes are intensely altered by dynamothermal metamorphism and metasomatic transformation. Throughout both orogenic regions quartz-feldspar pegmatites are numerous. These pegmatite bodies vary widely in size and shape, in internal texture, and in structural-geometric relations to their host rocks. The mineralogy of the West Greenlandic pegmatites is also variable, although quartz and feldspars are the dominant constituents of those studied. The validity of this statement depends upon the definition of the term pegmatite. If the term is applicable to all kinds of vein-, dike-, or sill-shaped bodies which are more coarse-grained and apparently younger than their host rocks, then the area contains numerous nonquartzofeldspathic pegmatites basically similar to the quartz-feldspar pegmatites in structure, texture, and relationship to the host rocks. Examples of nonquartzofeldspathic pegmatites are diopside-rich cross-cutting veins and conformable lenses of coarse diopside crystals in skarn rocks; small veinlets of garnet and biotite in metadiabases; veins of tourmaline in gneisses and schists (Pl. 8, fig. 4); concentrically zoned bodies of coarse-grained skarn minerals in limestones; clusters of very coarse-grained calcite-phlogopite bodies in smaller-grained marbles; and calcite-quartz veins in low-temperature greenstones (epidote amphibolites) on Hunde Iland. Some basic inclusions in gneisses have developed a coarse-grained concentric reaction structure not unlike some concentrically zoned pegmatite bodies.

Nonquartzofeldspathic pegmatites are less common than the pegmatites consisting chiefly of quartz and/or alkali feldspar or acidic plagioclase. However, the mechanisms of formation of the various kinds of bodies which can be called pegmatites in the broadest sense may well be basically the same, although the details of the formative mechanisms were different in the various cases, and the mobile elements which build the bodies were often dissimilar.

Acknowledgment

The field work in Greenland was sponsored by Greenland Geological Survey, Copenhagen. Most of the laboratory study was done at the University of Chicago. The manuscript was written during a period when the writer received a grant from The General Science Council of Norway, Oslo (Norges Almenvidenskaplige Forskningsråd).

Composition of the Pegmatites

The studied pegmatites range from practically pure quartz bodies to bodies consisting chiefly of feldspar; the latter is a rare type. Quartz-poor pegmatites are common in the fiord Söndre Isortoq where diopside-bearing gneisses formed by alteration of calcareous sediments are the prevailing host rocks. Pegmatites with intermediate quartz-feldspar ratio or slightly more quartz than feldspar are most abundant. Gradations from these types to pure quartz veins are much more common than are gradations from pure feldspar veins to quartz-feldspar pegmatites.

Some quartz veins poor in feldspar carry plagioclase exclusively (Ellitsgaard-Rasmussen, 1954, p. 62), whereas a very few carry potash feldspar exclusively, but most have both.

The types of feldspar present in the pegmatites with intermediate quartz-feldspar ratio vary considerably; some are pure plagioclase rocks, some contain chiefly potash feldspar, but most commonly plagioclase and potash feldspar occur together in the same pegmatites. Among the pegmatites containing but one kind of feldspar, plagioclase is considerably more common than potash feldspar. Plagioclase pegmatites are especially common in amphibolites and biotite schists, whereas alkali feldspar pegmatites may be found in granodioritic gneisses, although plagioclase commonly also occurs in pegmatites in such rocks.

Neither amazonite nor cleavelandite, so common in other pegmatitized regions (United States, Norway, Sweden), has been found in pegmatites in the investigated region.

Except for orthite, which is a typical but minor pegmatite mineral in certain districts in Greenland (Ameraliq fiord), so-called "rare" minerals are truly rare in the Greenlandic pegmatites. This holds for the rare-earth minerals as well as for the lithium micas and rare phosphates. Beryl is found only at a few localities. Tourmaline is common but not so abundant as,

for example, in the south Norwegian pegmatites. Other minerals which occur in the Greenlandic pegmatites are: biotite (very common), muscovite (generally rare, in some regions—granulite facies—not present), garnet (in some districts an exceedingly common pegmatite mineral), hornblende, diopside, hypersthene (confined to pegmatites in granulite-facies complexes), scapolite, sphene, epidote, cordierite, sillimanite, apatite, graphite, magnetite, pyrite, pyrrhotite, molybdenite, zircon, monazite, and an unidentified black pyroxene-like mineral which occurs in large quantities in feldspar-rich pegmatites between Söndre Isortoq and Alangua (Fig. 16).

These minerals constitute quanitatively a minor portion of the quartz-feldspar pegmatites, but they are important because they earmark the pegmatites and give information about the history of the pegmatites. Some of these minerals are confined to pegmatites in certain host rocks (*i.e.*, graphite, sillimanite, cordierite, diopside, and hornblende in pegmatites in host rocks rich in graphite, sillimanite, cordierite, diopside, and hornblende respectively), whereas other minerals occur in pegmatites confined to regions with certain grades of regional metamorphism. Orthite is an abundant pegmatite mineral in the region around Ameraliq fiord without any conspicuous relation to grade of metamorphism or type of host rock. However, the schists and common hybrid gneisses in this region generally contain more than the usual amount of orthite.

Distribution and Localization

Although pegmatites are scattered throughout the investigated part of West Greenland, they are not evenly distributed. Pegmatites are much more abundant in some regions than in others, and some types of pegmatites (characterized by given mineral content) are confined to certain regions.

Localization of pegmatites is (1) related to the grade of regional metamorphism in the area (this is true for both orogens), (2) determined by the mechanical properties of the host rocks, and (3) controlled by the character of the tectonic evolution of the regions (joints and faults generally localize pegmatites). The localization of special types of pegmatites (*i.e.*, characterized by a certain mineral composition, for example graphite-bearing pegmatites or pegmatites with diopside) seems to have been controlled by a combination of the mineral-chemical composition of the host rocks and the temperature and pressure affecting the complex at the time of emplacement of the pegmatites.

It is uncertain whether nearness to large granitic or granodioritic bodies affects the frequency of pegmatites in Greenland. Homogeneous massive granites or allied rocks are rare in the mapped area, yet pegmatites are very common.

In certain localities pegmatites may be very abundant, whereas in other localities with apparently the same lithological and structural characteristics pegmatites may be absent. This means that localization of pegmatites is somewhat erratic when measured in terms of readily observable geologic features.

The regional metamorphism within the mapped area has created three vast complexes of granulite-facies rocks; the remaining regions are chiefly occupied by amphibolite-facies rocks only locally altered to epidote-amphibolite facies in post-orogenic fault zones (Fig. 1). Schists and amphibolites (greenstones) belonging to epidote-amphibolite facies are situated on some little islands north of Egedesminde (Ellitsgaard-Rasmussen, 1954). Pegmatites, especially large ones, are relatively rare in all the granulite-facies areas. The amphibolite-facies areas (or the high epidote-amphibolite-facies areas) contain the largest number of pegmatites as well as the largest bodies of such rocks. The low-grade schists and amphibolites north of Egedesminde contain few pegmatites with intermediate quartz-feldspar content, although plagioclase-carrying quartz veins, pure quartz veins, and calcite-bearing quartz veins are found (Hunde Iland, Personal observation, 1948). As the spatial distribution of pegmatites by and large is related to the degree of regional metamorphism in the Greenlandic orogens, so, to some extent, is the mineralogical composition of the pegmatites. This mineral—grade-of-metamorphism relationship is especially conspicuous if the pegmatites of the granulite-facies regions are compared with those of the

amphibolite-facies regions. In the granulite-facies regions many pegmatites carry one or more of the following minerals: hypersthene, a dark-brownish-green hornblende, dirty-green or dark-gray antiperthitic plagioclase rich in small opaque inclusions, mesoperthitic alkali feldspar ($Ab_{50}Or_{50}$), a certain bluish quartz, manganese-poor garnets, sillimanite, and cordierite. In the amphibolite-facies regions hypersthene, brown hornblende, antiperthitic plagioclase, mesoperthitic alkali feldspar, and manganese-poor garnet fail to develop in the pegmatites (as well as in the country rocks, *see* Ramberg, 1948, p. 35–40). Instead the hornblende is of the common green variety, the garnet is rich in manganese, and the alkali feldspars do not show signs of extensive miscibility.

The localization of pegmatites is partly controlled by the mechanical properties of the host rocks. Basic rocks such as amphibolites, ultrabasics, and diopside skarn rocks are among the most common host rocks to the Greenlandic pegmatites. Pegmatites in such rocks are generally transgressive bodies, although conformable pegmatites do occur in schistose or layered varieties. These observations from Greenland agree with the findings of Andersen (1931) and others from Southern Norway where the majority of the pegmatites occur in amphibolites and other basic rocks. Andersen believed that dissimilar thermal expansion of basic and acidic rocks gave rise to cracks and fissures in the former; in this way he attempted to explain why pegmatites occurred more frequently in basic rocks than in acidic ones. It is true that cracks and ruptures are much more common in the basic rocks in regional metamorphic terrains than in acidic gneisses, and this is the reason why many pegmatites become localized in basic rocks. As to the reason for the rupturing of basic rocks, however, the author disagrees with Andersen. There is unmistakable evidence that basic rocks such as amphibolites, ultrabasics, diopside skarns, and anorthosites are much less ductile and susceptible to plastic recrystallization flowage than acidic quartz-alkali feldspar rocks (provided the latter are not too coarse-grained). Tension joints develop especially readily in basic rocks when intercalated with acidic gneisses or micaceous schists and exposed to tectonic forces. Such tension joints are favored sites for fixation of pegmatitic material. Basic rocks, owing to their tendency to rupture under tension rather than to flow plastically, function as sponges which extract mobile pegmatitic material wherever available. Amphibolite layers in the Greenlandic gneiss complexes are therefore very commonly full of crosscutting pegmatites in contrast to the adjacent quartz-feldspar gneisses in which only a few pegmatites generally occur (Figs. 2–6).

Many micaceous schists are also rich in pegmatites, much more so than the quartz-feldspar gneisses. In contrast to the crosscutting pegmatites which are typical of amphibolites and other basic rocks, most pegmatites in the mica schists are conformable, lenticular or pinching

FIGURE 2.—PEGMATITES IN AMPHIBOLITE
Mountain side with amphibolite layer (black) with crosscutting pegmatites in banded gneiss, Ikertoq fiord. Height of mountain about 300 m. Sketch after photo.

FIGURE 3.—PEGMATITES IN AMPHIBOLITE
Thick layer of amphibolite broken up and infiltrated by pegmatites, north side of Kobbefjord, Godthåb. Country rocks are banded and veined gneisses. Height of mountain about 300 m. Sketch after photo.

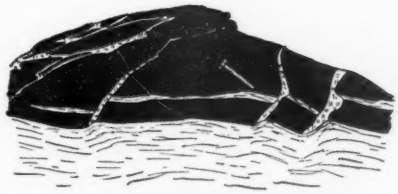


FIGURE 4.—PEGMATITES IN AMPHIBOLITE
Amphibolite inclusion with pegmatite-filled cracks in veined gneiss, Godthåb fiord. The amphibolite body builds a little hill in the terrain. Sketch after photo.

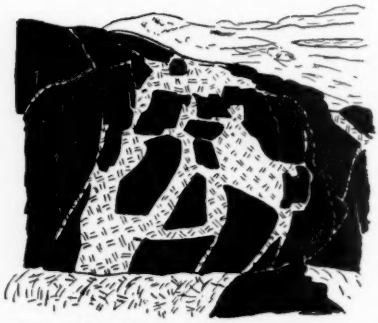


FIGURE 5.—PEGMATITES IN ULTRABASICS
Ultrabasic inclusion (black) broken and infiltrated by pegmatite material, in veined and banded gneisses, between Söndre Isortoq and Alangua. Knapsack on pegmatite in background shows size. The ultrabasic body contstitutes a little ridge. Sketch after photo.

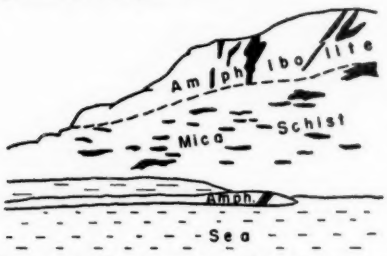


FIGURE 6.—PEGMATITES IN AMPHIBOLITE AND MICA SCHIST
Amphibolite with crosscutting pegmatites (black) resting on mica schist with conformable pegmatites, Laksebugt, Christianshåb. Top of mountain about 200 m high. Sketch after photo.

and swelling bodies as shown in Figure 6 and in Figures 1 and 2 of Plate 1. Many crosscutting pegmatites in mica schists exhibit ptygmatic folding, especially where the dikes are narrow.

Small conformable pegmatite veinlets are very abundant in some quartz-feldspar gneisses in the area studied, but large pegmatites are in general rarer in such rocks than in basic rocks and mica schists. Where the granodioritic gneiss is homogeneous and massive, such as in anticlinal cores and dome structures, pegmatites are especially scarce. Pegmatites in the quartz-feldspar gneisses are crosscutting as well as conformable, but crosscutting structures are more common than conformable ones among the larger pegmatite bodies. Most small pegmatite veinlets in quartz-feldspar gneisses, however, are conformable schlieren.

Joints and faults which localize pegmatites in Greenland traverse all kinds of rocks in the area. The statistical distribution of these kinds of ruptures, however, is controlled by the rock type. The relative tendencies of rocks to rupture under tensional stress are generally as follows: olivine-pyroxene rocks are most brittle (ultrabasics, diopside skarn rocks) and show little ability to yield plastically, various types of amphibolites are somewhat less brittle, and fine- and medium-grained quartz-alkali feldspar rocks are the least brittle and exhibit a pronounced tendency to yield by plastic flowage. Mica schists are even more plastic than quartz-feldspar gneisses. Coarse-grained quartz-feldspar bodies such as pegmatites appear more brittle or competent than mica schists and small-grained gneisses. This empiric observation explains why transgressive pegmatites are much more common on basic rocks than in schists and gneisses but does not contradict the fact that crosscutting pegmatites do occur in gneisses and mica schists, and that some pegmatites may traverse intercalated layers of basic rocks, quartz-feldspar gneisses, and mica

PLATE 1.—PEGMATITES

1, 2. Medium-grained quartz-feldspar pegmatites in mica schist, Ikamiut. Photograph, H. R., 1948

3. Folded pegmatites, in part showing pinch-and-swell structures in veined gneiss, Ameraliq fiord. Photograph, H. R., 1951

4, 5. Examples on pegmatites replacing fine-grained biotite-bearing quartz-feldspar dikes, Godthåb fiord, Photographs, H. R., 1949 ahd 1951

PLATE 2.—PEGMATITES

1. Series of uniformly thick and straight pegmatite dikes cutting amphibolite, Godthåb fiord. Photograph, H. R., 1951

2. Ptygmatically folded pegmatite veins in gneiss, Godthåb fiord. Photograph, M. L. Ramberg, 1951

3. Intensively folded pegmatite in veined gneiss, Godthåb fiord. Photograph, H. R., 1951

4. Ptygmatically folded pegmatites cut by younger straight pegmatite, Kangersuneq fiord south of Christianshåb. Photograph, H. R., 1948

FIGURE 3

FIGURE 5

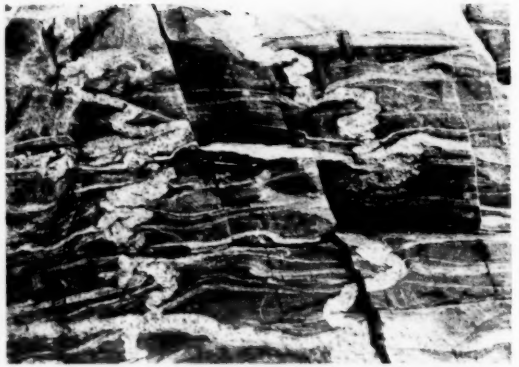
FIGURE 1

FIGURE 2

FIGURE 4

PEGMATITES

FIGURE 1



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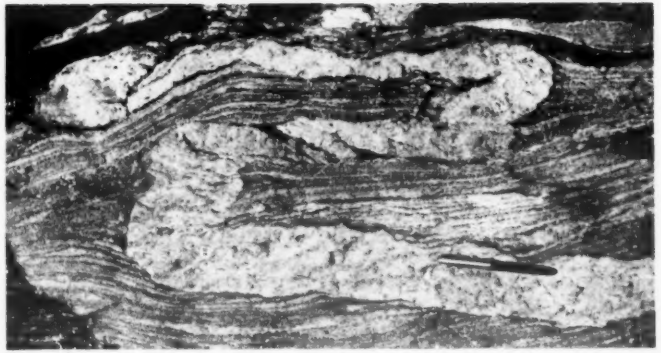

FIGURE 3

FIGURE 4

PEGMATITES

FIGURE 1

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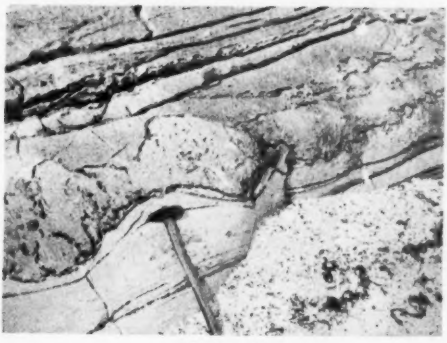


FIGURE 4

FIGURE 5

PEGMATITES

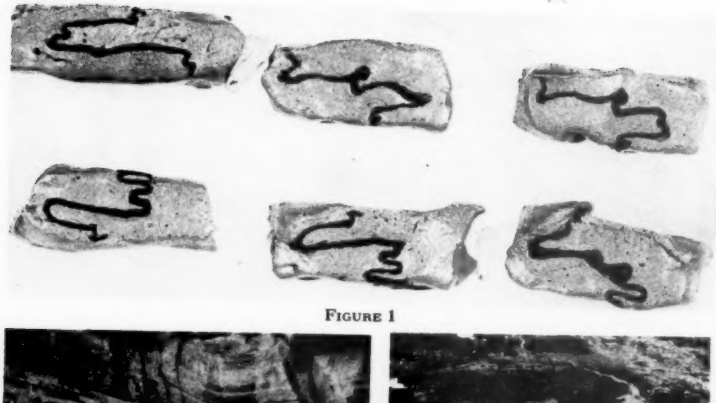
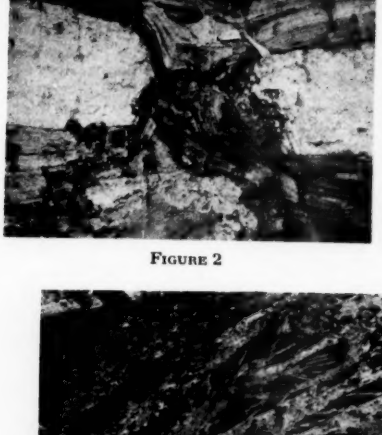


FIGURE 1



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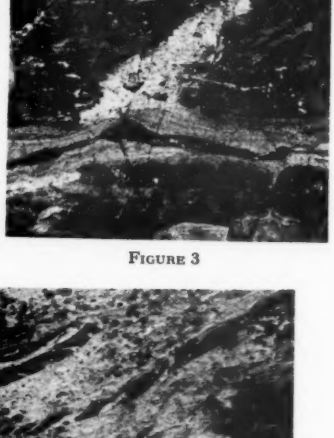

FIGURE 3

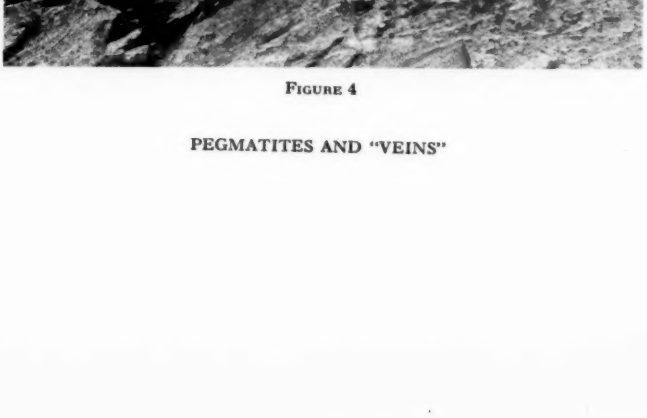

FIGURE 4

PEGMATITES AND "VEINS"

schists. The tendency to rupture under tensional stress or shear depends upon the magnitude of the stress, the length of application of the stress, pressure and temperature conditions, whether the rock is "dry" or "wet", the presence of substances which can act as catalysts, and other factors.

More than one system of faults and joints is developed in many places in the area studied. In such cases pegmatites do not distribute themselves evenly among the fracture systems but may be localized in one system or in a limited number of such systems. Along the north shore of Fiskefjord, for example, there are vertical cross-joints and sets of diagonal fractures. In this district most pegmatites follow a diagonal fracture system dipping about 30° E.NE. along which faulting also has taken place. The fracture system along which faulting took place developed during the dynamothermal metamorphism because newly formed minerals in the blastomylonites in this fault system are not low-temperature minerals but conform usually to the regional metamorphism of the country rocks. The other joint systems, in particular the cross-joints, probably did not develop as real breaks until after the complex was elevated—*i.e.*, after the dynamothermal metamorphism and thus too late for pegmatite growth.

Pegmatites are localized along certain fine-grained quartz-feldspar dikes which occur abundantly in the regions around Godthåb fiord and around Holsteinsborg settlement. Many of these fine-grained dikes, which are thoroughly altered during the metamorphism (if not formed by metamorphic-metasomatic action), are replaced by synmetamorphic pegmatites along the margins, in the central portions, or in a more irregular fashion as shown in Figures 4 and 5 of Plate 1.

Pegmatitic matter also concentrates around basic inclusions in the schists and gneisses (Pl. 3, fig. 1; Pl. 6, figs. 3, 4). It is believed that these pegmatites grew in the pressure shadows of the rigid inclusions or that quartz and feldspar replaced the ferromagnesian minerals in local spots of maximum tensile stress in originally homogeneous and continuous layers of basic rocks. Tension generally develops in a strong or competent basic layer enclosed in competent rocks if the complexes are exposed to one- or two-dimensional compression. Rotation of the inclusions causes pegmatite minerals to grow around the inclusions in a snowball-like fashion.

Size and Shape of Pegmatites

General

The investigated pegmatites range in size from veinlets only millimeters thick or lenses a few cubic centimeters in volume to bodies

PLATE 3.—PEGMATITES

1. Pegmatitic quartz-feldspar material concentrating around inclusions of diopside skarn in quartz-feldspar gneiss. Söndre Isortoq. Photograph, M. L. Ramberg, 1949
2. Pinch-and-swell pegmatites in banded and veined gneiss, Godthåb fiord. Photograph, H. R., 1951
3. Pinch-and-swell pegmatite in veined gneiss, Faeringerhavn. Reproduced from Kodachrome, H. R., 1948
4. Pinch-and-swell structure of monomineralic diopside "pegmatite" in dolomite marble, Almenningöy, Tröndelag, Norway. Photograph, H. R., 1940
5. Extreme pinch-and-swell structure of pegmatites in mica schist, Black Hills, South Dakota. Photograph, H. R., 1949

PLATE 4.—PEGMATITES AND "VEINS"

1. Sections of experimental ptygmatically folded "veins" (relatively competent plastecine) in incompetent putty. The plastecine "veins" were originally straight vertical sheets in the putty. Compressive stress was applied in the vertical direction. Each section about 4 cm long.
2. Boudins of medium-grained quartz-feldspar pegmatite in micaceous quartzite, west of Atlantic City, Wyoming. Photograph, M. L. Ramberg, 1949
3. Wedge-shaped quartz-feldspar pegmatite extending from sedimentogeneous gneiss into layer of diopside skarn, Amitssuarsoralaq. Photograph, H. R., 1949
4. Folded and fractured crosscutting amphibolite dike in gneiss, Ameraliq fiord. Length of photo about 5 m. Photograph, M. L. Ramberg 1951

which may reach a thickness of up to 30 m and a length of several hundred m. The abundance of pegmatites is approximately inversely proportional to their size; small pegmatites are overwhelmingly common, the larger pegmatites are rare. Some gneisses contain up to 50 per cent by volume of small conformable quartz-feldspar pegmatites.

The pegmatites vary much in their shape, in striking contrast to the straight and uniformly thick diabase dikes that swarm through the southern portion of the mapped area.

Sheet-Shaped Pegmatites

Some pegmatites occur as sheets which either are crosscutting or conformable (Pl. 2, fig. 1). Unlike most unaltered diabase dikes, however, most sheet-shaped pegmatites are not uniform in thickness but pinch and swell to various degrees, particularly conformable ones (Pl. 3, fig. 2). The sheet-shaped pegmatites are rarely straight but tend to be folded.

The extension of such sheet-shaped pegmatite bodies is everywhere very limited. No single body found extends more than a few hundred meters along the well-exposed erosion surface. The pegmatites may thin out gradually or end abruptly at the border between two contrasted rock, such as between amphibolite and mica schist or quartz-feldspar gneiss. Crosscutting pegmatites may penetrate the amphibolite but end abruptly at the border toward schist or gneiss (Pl. 7, fig. 4).

Ptygmatically Folded Pegmatites

Most sheet-shaped pegmatites are folded, the conformable bodies following the folds in the adjacent rocks and the crosscutting bodies being ptygmatically folded (Pl. 2, figs. 2, 3, 4). If several generations of crosscutting pegmatites occur the oldest dikes are generally ptygmatically folded whereas the younger dikes may be straight (Pl. 2, fig. 4). Many theories have been proposed to explain ptygmatic structure of veins, but for the ptygmatically shaped pegmatites in Greenland no explanations other than a simple one- or two-dimensional compression of the plastic (incompetent) host rocks and a consequent folding of the less plastic (more competent) pegmatite dikes is consistent with observed facts. In some cases the strain of the pegmatites under the process of ptygmatic folding has resulted in faintly developed schistosity parallel to the shear planes (Pl. 2, fig. 3). In many instances the detailed structure in the host rock adjacent to the sharp bends in the pegmatites shows that the pegmatite was folded as a relatively competent body which in part controlled the structural evolution of the near-by country rock (Pl. 2, fig. 3). Similar features are developed along folded crosscutting amphibolite bands in Ameraliq fiord (Pl. 4, fig. 4). In the host rock around ptygmatically folded pegmatites and quartz veins at several localities in Scandinavia structural indications of secondary or syngenetic folding of the competent crosscutting veins are widely exhibited (unpublished observations). Godfrey (1954, p. 375–387) has shown that ptygmatic veinlets in schists from Yukon have been folded simultaneously with or following their emplacement.

Post- or synemplacement folding of the ptygmatic pegmatite veins in Greenland is not always clearly demonstrated by macroscopic features; however, in general the following observations indicate that ptygmatic folds are formed in the crystalline state as a result of compression of the plastic enclosing rock perpendicular to its planar or linear structure: (1) The fold axes of the ptygmatic veins are approximately parallel to the plane of schistosity (or the lineation) in the host rock. (2) The general direction of ptygmatically folded veins is inclined to the general schistosity plane in the host rock, although a narrow contact zone of secondary conformable schistosity may develop close to the folded vein. (3) In the investigated part of West Greenland there are independent proofs of intensive compression of the gneisses and schists in a direction perpendicular to the schistosity plane. The compression was made possible by a compensating plastic expansion in one or two dimensions parallel to the linear or planar structures. The expansion took place preferably in the most incompetent or ductile rock layers in the gneissic complexes, so that tensional stress and consequent jointing and formation of boudinage structures affected the less ductile or more

competent layers (the amphibolites and other basic rocks). Pinch-and-swell structure in pegmatites develops as a response to tensional stress along conformable pegmatites because of drag owing to plastic flowage in adjacent incompetent host schists or gneisses. Ptygmatic folds and pinch-and-swell structure are associated in many localities. Indeed, conformable portions of a ptygmatically folded pegmatite may develop pinch-and-swell structure as shown on Figure 3 of Plate 1. (4) A structure very similar to ptygmatically folded veins may develop when transverse schistosity is impressed on sediments in which competent and incompetent layers alternate. The competent layers become ptygmatically folded, and the adjacent incompetent layers develop a transverse recrystallization schistosity. This is observed in the sparagmite formation in Norway (Fig. 7). (5) Experiments show that ptygmatic folds readily develop by the mechanism explained in (4) (Pl. 4, fig. 1).

Most of the ptygmatically folded pegmatites are not more than 10–25 cm thick. Only rarely are larger pegmatites folded ptygmatically, such, as the several-meters-thick transgressive pegmatite lenses in veined gneiss on the east side of the island Sadelen, Godthåb fiord.

Pinch-and-Swell Structure

Many conformable pegmatites in mica schists and quartz-feldspar gneisses have pinch-and-swell structure (Pl. 3, figs. 2, 3, 5), but few of those in amphibolites and other basic rocks do. The pinch-and-swell structure is most readily explained if host rock and pegmatite were solid crystalline aggregates when this structure evolved. Pinch-and-swell structure in conformable pegmatites and quartz veins and boundinage structure in amphibolites or other basic layers in gneisses are basically similar in origin, as both structures are created by yield of relatively competent bodies under tension. The narrow portions of a pinch-and-swell pegmatite are not caused by pinching owing to localized compressional stress perpendicular to the sides of the pegmatite, but rather to local thinning or "necking down" owing to stretching parallel to a tensional stress acting along the pegmatite bodies. This is consistent with the relative thickness of the layers in the adjacent schist or gneiss; the compressional stress must have been relatively small at the narrow portions of the pegmatite because the adjacent schist or gneiss is relatively thick-layered here, and relatively high at the thick portions where the layers in the adjacent gneiss are relatively thin (Pl. 3, fig. 3). A rhythmic thinning of pinch-and-swell pegmatites as caused by literal pinching would not have occurred at places where the compressional stress was at a local

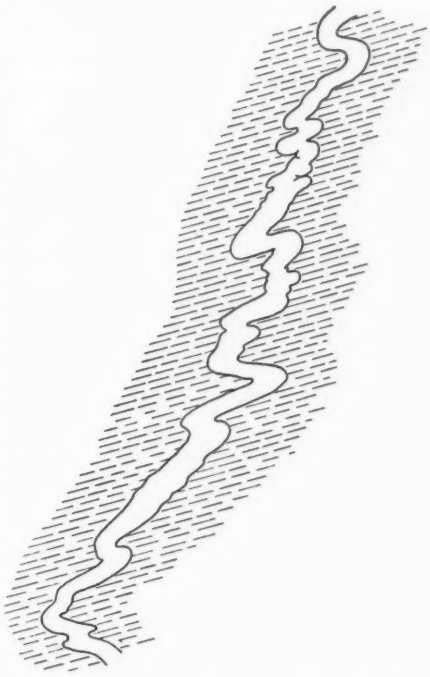

FIGURE 7.—FOLDED QUARTZITE LAYER IN MICA SCHIST

Ptygmatically folded quartzite band in mica schist. Secondary transverse schistosity has developed in the mica schist which originally was clay-rich layers in a varied sediment. The incompetent micaceous layers have been shortened considerably in a direction parallel to the original layering or perpendicular to the present schistosity. In response to the shortening the competent sandstone (quartzite) layer developed "ptygmatic" folds. Near Oppdal, Norway. Sketch of hand specimen. Length of quartzite band about half a meter.

minimum. The explanation of ptygmatic folding was based on the assumption that pegmatite bodies are competent relative to quartz-feldspar gneisses and mica schists. This assumption is supported by the fact that pegmatites are coarse-grained and therefore will not yield readily by recrystallization flowage under regional metamorphism. Based on the same assumption, the pinch-and-swell structure is readily explained in principle as follows: In response to compressional stress the plastic quartz-feldspar gneiss or mica schist adjacent to the conformable pegmatite yields by flowage parallel to the planar or linear structure. This means that every small area parallel to the plane of schistosity in the host schist becomes enlarged in one or two dimensions in the course of the flowage. Along the pegmatite boundaries the lateral expansion of the host schist results, because of drag coupling, in tensional stress in one or two dimensions in the pegmatite parallel to the boundary surface. Because the pegmatite is less susceptible to yield by recrystallization flowage than are the enclosing rocks, it will not yield homogeneously along its entire length; instead the stretching is apt to be localized at sites along the pegmatite body. The initiation of yield at a given site is apparently due to some original heterogeneity, but once the yield has started at a locality the diminished cross section concentrates the tensional stress in this place, and stretching will continue here. This is exactly what happens when a metal rod is subjected to tension above its yield point: instead of occurring uniformly along the rod, the yield concentrates at a small portion of the rod thus creating one or occasionally more necked-down regions. In contrast to a stretched metal rod which usually yields at but one locality the pinch-and-swell pegmatites have yielded at numerous places with rather regular spacing along the bodies. Such a difference in behavior might well be expected because the yield at one local region in a pegmatite releases the tensional stress only within a limited portion of the pegmatite around the site of yield. Some distance from the yielding locality the tensional stress in the pegmatite as caused by the one- or two-dimensional expansion of the adjacent rock is unaffected by the local release, and the tensional stresses will initiate a new locality of stretching.

This process in pegmatites is very similar to the evolution of boudinage structures of basic layers in gneisses and micaceous schists (Pl. 5, fig. 2). Owing to the very low ductility of many basic rocks they commonly respond to the tensional stress by rupture rather than rhythmic regions of stretching and thinning, although such a behavior may also be exhibited by basic layers. Pegmatites enclosed in micaceous quartzites may develop true boudinage structure as shown in Figure 2 of Plate 4. In the region around Atlantic City, Wyoming, some pegmatites occur in micaceous quartzites. The thickest pegmatites were conformable sills or flat lenses. Thinner pegmatites from 10- to 30-cm thick were ruptured by tension joints to boudins separated by plastic flowage in the host schist. (A little quartz and biotite have formed in the pressure shadows between the boudins.) Still thinner pegmatites (1–5 cm thick) developed pinch-and-swell structure, but many of the thinnest pegmatite veins were ptygmatically folded. The varied behavior of the Atlantic City pegmatites is probably related to their thickness and corresponding textural coarseness, because all contain the same minerals and occur in the same schist and very close together. Many of the thickest bodies were too strong to yield by rupture or stretching under the tensional stress caused by the plastic flow of the host schist. Thinner pegmatites could not resist the tension and broke by rupture. Still thinner and smaller-grained pegmatites yielded to the tensional stress partly by plastic stretching and thus developed pinch-and-swell structure. All these pegmatites were strong enough to create a secondary conformity in the host schist and to rotate without break. Only the thinnest pegmatites were weak or plastic enough to develop ptygmatic folds.

Experiments have proved that pinch-and-swell structure can develop from the mechanism discussed (Ramberg, 1955). The writer believes it is physically impossible that the kind of pinch-and-swell structure shown in Plate 3 could develop by injection of the pegmatite as a melt. To affect the structure of the adjacent rock as these pegmatite bodies

have done, the bodies must have been able to support local stresses (pressure gradients) over a long period of time. Indeed the pegmatite bodies must have been stronger than the enclosing solid rocks which have been deformed. It seems incredible that a melt should be more rigid than a solid, in particular in this case where the chemical composition of both is so similar. More probably, the pegmatites were solid bodies during the evolution of the pinch-and-swell structure, and the greater competency (less plasticity) of the pegmatites is related to their coarser grain size which makes them less apt to yield by recrystallization flowage. Pinch-and-swell structure is common in many types of conformable veins which could not have been injected as melts. Figure 4 of Plate 3, for example, shows pinching and swelling veins of pure coarse-grained monomineralic diopside in dolomite marble. The texture of this body and its shape are identical to those of many pegmatites, but this diopside skarn "vein" was certainly not injected as a molten mass.

This interpretation of ptygmatic and pinch-and-swell structures conforms with the observation that pegmatites in mica schists in Greenland generally are conformable bodies with lenticular or pinch-and-swell shape. The conformable structure in these plastic host rocks to a great extent must be secondary; some of the pegmatites started out in the form of transgressive bodies which gradually became rotated and made conformable to the tectonic flow structure of the schist. If the schist is much more plastic than the pegmatites, the latter need not be conspicuously mylonitized.

It should be emphasized that it is not believed that the pinch-and-swell structure always formed after the pegmatite was emplaced; rather, the pegmatites grew as solid bodies such that the pinch-and-swell structure developed contemporaneously with the growth of the pegmatite body. The same is true of ptygmatically folded pegmatites; this folding may well be syngenetic.

Pegmatites with Irregular Shape

Many pegmatites are irregular in shape. Some are amoeba-shaped or irregularly branching bodies (Figs. 5 and 8), and labyrinthic networks of pegmatitic material penetrates amphibolites or other fragile rocks. Conformable branches of such irregular pegmatite masses may be pinching and swelling, and crosscutting portions may be ptygmatically folded.

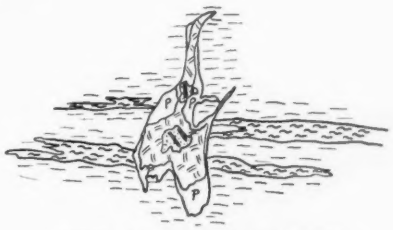

FIGURE 8.—PEGMATITE IN GNEISS

Branching and concentrically zoned pegmatite in heterogeneous gneiss, north side of outlet of Söndre Isortoq. P: border zone of fine-grained plagioclase and quartz; cross-hatched: intermediate zone of quartz, potash feldspar, and a little plagioclase; dotted: quartz core with biotite flakes (black). Conformable bodies with wavy signature are old deformed pegmatites. Zoned pegmatite is about 4 feet across.

Wedge-shaped pegmatites are not uncommon in the area. This shape often develops in crosscutting pegmatites which transect a certain rock layer in banded complexes and end abruptly at the boundary between the host layer and an adjacent dissimilar rock as shown in Figure 3 of Plate 4.

Pegmatite Fillings Between Boudins

Some of the most widespread pegmatites in the area occur as fillings between boudins of broken basic layers in mica schists and quartz-feldspar gneisses. Such pegmatites exhibit a distorted rectangular shape on exposed surfaces (Pl. 7, fig. 4; Pl. 10, fig. 2). An hourglass shape is also common (Pl. 5, fig. 2). In three dimensions the shape of such pegmatites may be complicated because some boudinized amphibolite layers are fractured by two sets of simultaneous joints which are perpendicular to the planar structure and intersect at an angle around 90°. If the pegmatitic material fills the two sets of joints in an extensive boudinized and folded amphibolite layer, the pegmatite body is shaped as a complicated and extensive network. In other cases boudinage

is caused by but one set of cross-joints which are perpendicular to the lineation of the host schist or gneiss. Under such circumstances the pegmatites which fill the space between boudins are long, ruler-shaped bodies.

The localization and fixation of the pegmatitic bodies in the interstices between boudins give valuable information as to the mechanisms of formation of pegmatites in general. With the theory of pinch-and-swell structure in mind it is pertinent to discuss the pegmatites shown in Figure 2 of Plate 5. The picture shows that there must have been a considerable stretching of the pegmatites between the boudins under or after the emplacement of the pegmatitic material. This stretching, which took place in the crystalline state, was due to an elongation of the host schist parallel to its planar structure and a consequent separation of the individual boudins of the ruptured amphibolite layer. The pegmatites between the boudins therefore became considerably stretched parallel to the schistosity of the adjacent schist and thinned perpendicular to the schistosity. Most likely the pegmatite grew in the course of the stretching. This is very similar to the mechanism which creates necked-down regions in pinch-and-swell pegmatites.

In Figure 1 of Plate 5 it is significant that the planar structure of the schist bends around the border portion of the pegmatite thus giving the impression that the border portions of the pegmatites (upper contact in the figure) were crystalline and adhered to the ends of the adjacent amphibolite boudins when the central portion of the pegmatite was emplaced. This indicates that the joints in which the pegmatite formed opened slowly and in pace with the growth of the pegmatite. Thus the present pegmatite-occupied space was probably never a void or fluid-filled space; this is also suggested by the fact that unfilled or partly filled openings between boudins have never been found in Greenland. If the boudins were separated faster than the pegmatitic minerals became emplaced, one should expect now and then to find half-filled openings between boudins. Other phenomena also make it hard to believe that the spaces between boudins were ever empty in the intensively metamorphosed Greenlandic gneisses. The very cause of the formation of boudinage structure is the plastic flowage of incompetent ductile layers adjacent to competent brittle layers. Therefore, if pegmatitic or other vein material were not available to enter the expanding joints, then the adjacent incompetent schist or gneiss would be plastically injected into the joints and would fill them. In some places this has happened in Greenland, but by far most interstices between boudins contain appreciable amounts of vein material although some plastic injection of the adjacent incompetent rock has almost

PLATE 5.—PEGMATITES

1. Pegmatite between two amphibolite boudins (upper and lower left side of the picture) in mica schist (right side of picture). Note intensive replacement of portion of schist which has crept into region between boudins. Cluster of black mineral in pegmatite is tourmaline. Laksebugt, north of Christianshåb. Photograph, H. R., 1948

2. Stretched and "necked down" hour-glass shaped quartz pegmatites between amphibolite boudins in mica schist, Godthåb fiord. Photograph, H. R., 1951

3. Pegmatite with pure quartz core and quartz-feldspar boarder zone in veined gneiss, Ameraliq fiord. Photograph H. R., 1951

PLATE 6.—PEGMATITES

1. Gradational contact between pegmatite rich in potash feldspar and granite gneiss (right-hand side of picture). Note that banded country rock on left-hand side is crumpled along contact with the pegmatite. Proximity of Christianshåb. Reproduced from Kodachrome, H. R., 1948

2. Coarse potash feldspar, quartz, and biotite replacing granitic gneisses (early stage of evolution of replacement pegmatite). Proximity of Christianshåb. Reproduced from Kodachrome, H. R., 1947

3. Pegmatitic material concentrating around amphibolite inclusions in gneiss, Ameraliq fiord. Photograph, H. R., 1951

4. Layer of amphibolite densely infiltrated by pegmatitic material. Ameraliq fiord. Photograph, H. R., 1951

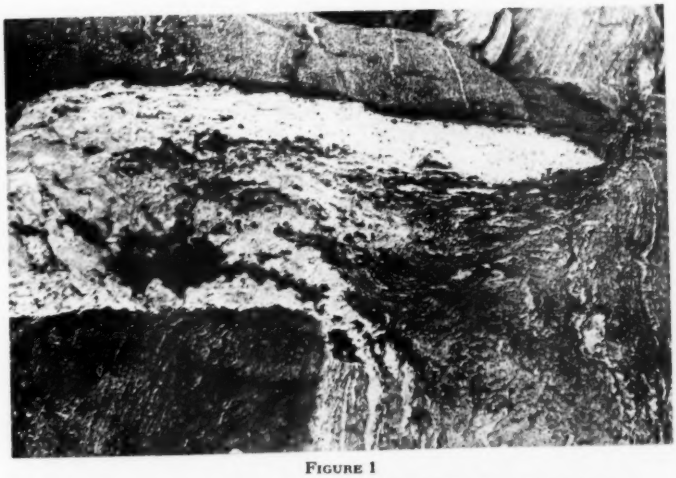

FIGURE 1

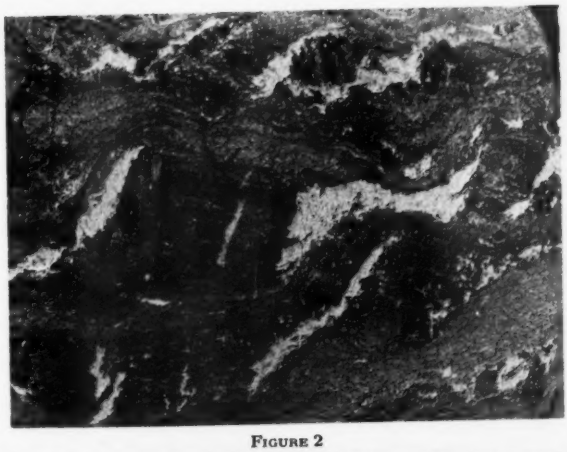

FIGURE 2

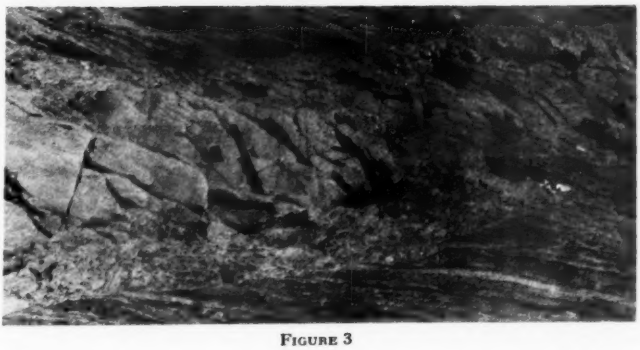

FIGURE 3

PEGMATITES

FIGURE 1

FIGURE 2

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FIGURE 4

PEGMATITES

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PEGMATITES

FIGURE 2

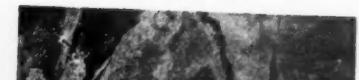

FIGURE 4

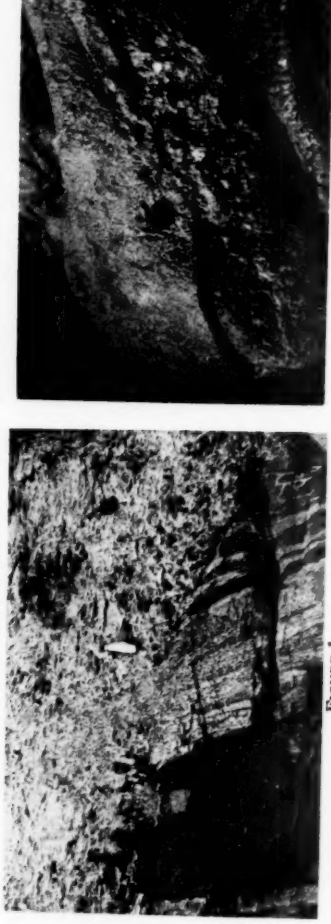

FIGURE 1

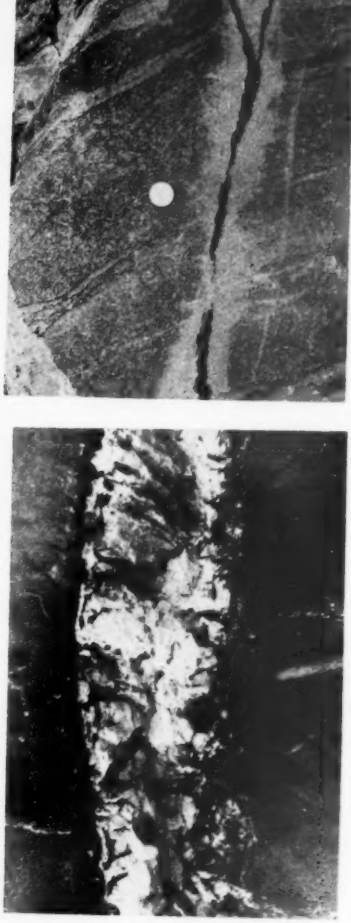

FIGURE 3

PEGMATITES

invariably taken place. Thus the evidences from the studied area suggest a strong synchronization between the spatial dilation of the boudin-producing ruptures and the filling of these openings by pegmatites or quartz veins.

In Figure 3 of Plate 7 the *S*-shaped gneiss bridge which crosses the pegmatite body proves that the dilation of the joint took place by a motion of the country rock inclined to the joint (indicated by the arrows on the figure). The question is when did this motion occur relative to the emplacement of the pegmatite. Since the gneiss bridge is plastically folded and considerably sheared in its central portions, it is exceedingly unlikely that the fissure was empty when the dilation occurred. If the thin gneiss bridge were not supported by some material with mechanical properties similar to those of the gneiss itself one should expect that the bridge would have fractured rather than been plastically deformed. It seems highly improbable that the supporting material could have been a liquid with the high fluidity usually attributed to pegmatite magmas. If the folded gneiss bridge were supported by a fluid, one should expect similar structure in some of the numerous diabases in the area, but none has ever been found. The writer is not familiar with comparable structures described from diabase dikes or other unquestionable magmatic dikes from other parts of the earth. It is therefore believed that the gneiss bridge across the pegmatite was supported by surrounding crystalline aggregate during every stage of its evolution, and consequently that the pegmatite mineral accumulated in pace with the dilation of the fissure.

Pegmatites Wrapped Around Basic Inclusions

Pegmatitic quartz-feldspar material is commonly associated in gneisses with diopside skarn, ultrabasite inclusions, or other basic inclusions around which it wraps like an envelope. This type of pegmatite occurrence is explainable by dissimilar mechanical properties of the basic inclusion and the enclosing gneiss, the former yielding by rupture, the latter by plastic flowage. Pegmatitic minerals will grow in pressure shadows in lee of the rigid inclusions in much the same fashion that quartz grows in pressure shadows of garnet prophyroblasts. If the inclusions become rotated in the gneiss, the lee or pressure-shadow side of the inclusion will be displaced along their surfaces, and pegmatite minerals can continue to grow around the entire inclusion in snowball fashion. The coarse-grained pegmatite mineral aggregate is strong and competent so that it will not necessarily be squeezed out under the rotation of the inclusions. Most inclusions with wrapped-around pegmatite material are concentrated in a zone parallel to the schistosity, thus indicating that the inclusions are remnants of

PLATE 7.—PEGMATITES

1, 2. Gradational contacts between micaceous gneiss and conformable pegmatites, two localities in Godthåb fiord. Photograph, H. R., 1951

3. Crosscutting pegmatite with distorted S-shaped gneiss bridge across, Godthåb fiord. Photograph, H. R., 1949

4. Pegmatite with roughly rectangular shape situated between amphibolite boudins in hypersthene gneiss, Tovqussaq. Photograph, M. L. Ramberg, 1949

PLATE 8.—PEGMATITES

1. Details of contact between pegmatite and banded gneiss. Note how a light band in the gneiss continues as a diffuse fine-grained region into coarse-grained pegmatite body (left of knife). A little south of Söndre Isortoq. Photograph, H. R., 1949

2. A 12-cm crosscutting pegmatite, *p*, showing nondilation geometry relative to country rock. Dark-gray band, *a*, in the enclosing gneiss is a prepegmatite crosscutting hornblende-rich "dike"; *b* is biotitized remnants of *a* in the pegmatite. South of Netsilik between Kangeq and Fiskefjord. Photograph, H. R., 1949

3. Potash feldspar-plagioclase-quartz-biotite pegmatite with coarse biotite flakes along the contact toward amphibolite, proximity of Christianshåb. Reproduced from Kodachrome, H. R., 1948

4. Tourmaline vein with bleached external contact zone (extraction zone) in biotite- and hornblende-bearing gneiss, Godthåb fiord. Photograph, H. R., 1951

continuous basic layers which have been partly replaced irregularly by coarse quartz and feldspar and partly invaded by these minerals in tension fractures. Figure 4 of Plate 6 represents an initial stage of the process.

FIGURE 9.—PEGMATITE IN GNEISS

Crosscutting pegmatite in banded gneiss. Bands with fine hatching are mica-rich layers. Coarse horizontal hatching indicates amphibolite bands. Especially large feldspar crystals in pegmatite are black. Some prophyroblasts are shared between gneiss and pegmatite along the border, and the pegmatite is merely an assemblage of feldspar porphyroblasts where it crosses micaceous layers in the gneiss. Thickness of pegmatite about 8 cm. Amitssuarsoralaq, south of Söndre Isortoq.

STRUCTURAL RELATIONS BETWEEN PEGMATITE AND COUNTRY ROCK

The contact between pegmatite and enclosing rock may be macroscopically sharp (Pl. 11, fig. 4) or diffuse and transitional (Pl. 7, figs. 1, 2; Pl. 6, figs. 1, 2). Even where the contact is macroscopically sharp, however, the microscope usually reveals a diffuse contact with mineral grains shared between pegmatite and host rock. This is generally observable in the field because the grains of most shared minerals are big enough to be detected by the naked eye.

That pegmatites and host rocks are intimately intergrown or welded together is also indicated by the weathered rock which generally does not show a discontinuous break along the contacts. This is in sharp contrast to the weathered morphology along most fresh diabase dikes in the area.

Many contacts between pegmatite and host rock are diffuse, showing that the pegmatite replaced the host rock. This is most commonly developed in transgressive pegmatites (Pl. 6, figs. 1, 2; Pl. 8, fig. 1). In conformable pegmatites gradation toward enclosing rocks is exhibited by banded or schlieren structure in

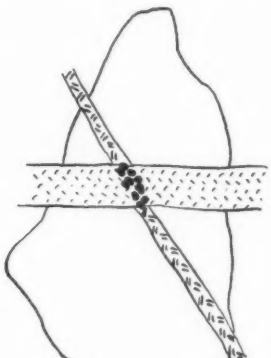

FIGURE 10.—INCLUSION WITH PEGMATITE IN GNEISS

Ultrabasic inclusion in gneiss crossed by gray "aplite" dike (horizontal) and plagioclase pegmatite, island between Kangeq and Fiskefjord. The gray "aplite" dike seems to cut across the pegmatite, yet the pegmatite continues through the "aplite" as a shadowy zone of coarse porphyroblasts. Thickness of "aplite" about 12 cm.

the pegmatite (Pl. 7, figs. 1, 2). Some pegmatites are merely clusters of feldspar prophyroblasts in gneisses or schists, and all gradations can be found between single prophyroblasts and larger pegmatite bodies.

Dissimilar layers in banded gneisses may respond dissimilarly to replacement pegmatization. Some of the most resistant bands project into or through pegmatites as ghostly remnants, and in places a pegmatite with a macroscopically sharp boarder toward some rock layers penetrates other layers as diffuse ghostly shadows (Figs. 9, 10; Pl. 8, fig. 1).

Crumpling and plastic distortion of schists and gneisses adjacent to pegmatite bodies are common (Pl. 6, fig. 1). Such structures around pegmatite bodies have been taken as indications of forcefully injected pegmatite magma. It seems, however, unbelievable that a fluid melt could cause intense plastic deformation in solid rocks. Rather, the host-rock distortions

are in part due to the force of crystallization of the pegmatite bodies or, more commonly, are a result of tectonic strain during and after the emplacement of the pegmatites.

Many of the studied pegmatites are surrounded by external contact areolae of various kinds. Some of these contact areolae are results of readily understandable chemical interactions between pegmatite and minerals of the host rocks, such as the biotite zones around pegmatites in amphibolites (Pl. 8, fig. 3), biotite and/or hornblende zones around pegmatites in ultrabasic rocks, skarn minerals around pegmatites in limestones, and zones rich in hypersthene around some pegmatites in amphibolites in granulite-facies areas. Other external contact zones around the Greenlandic pegmatites are characterized by a relative enrichment of one or more of the host-rock minerals. Enrichment of hornblende around pegmatites in hornblende gneisses is especially typical of many small pegmatite veins or dikes. The quartz-feldspar components of the host rock have apparently migrated into the pegmatite thereby causing a relative enrichment of hornblende in the region close to the pegmatite. This common kind of external contact areolae has tentatively been termed extraction areola (zones) by the writer in order to emphasize the manner of formation. Extreme cases of extraction areolae are found around tourmaline veins in micaceous paragneisses in Godthåb fiord. These veins are surrounded by a bleached zone of quartz and feldspar from which all traces of ferromagnesian minerals (essentially biotite) have been extracted (Pl. 8, fig. 4). The iron and magnesium from the ferromagnesian minerals of the host rock (chiefly biotite) have been utilized in the building of tourmaline in the vein. The tourmaline does not fill a dilated fissure but replaces the host rock along an unopened joint. Pegmatites with external contact zones rich in biotite are found in biotite gneisses or schists.

Commonly the enriched minerals in the extraction areolae are more coarse-grained than in the unaffected portion of the host rocks; this is typically found in hornblende-rich zones around pegmatites in amphibolites or hornblende gneisses and also in biotite-rich zones around pegmatites in micaceous gneisses or schists. This must mean that the enrichment of certain minerals in the extraction zones is accompanied by some recrystallization of the relatively concentrated mineral, or by introduction of the constituents of this mineral to the external contact zone. The latter process is explainable in terms of the concept of basic front. When a quartz-feldspar pegmatite body makes room in, for example, an amphibolite by means of replacement some of the elements constituting hornblende are rendered superfluous because they are not completely utilized in building ferromagnesian pegmatite minerals. These elements therefore are forced to migrate away from the growing pegmatite petroblast. The superfluous elements may become fixed just outside the pegmatite by simple "precipitation" on pre-existing hornblende grains. However, for the hornblende grains to enlarge, space must be provided or the superfluous elements coming from the site of pegmatitization cannot be fixed in the external contact zone. Space for enlargement of the hornblende grains in the contact areolae could be provided by removal of quartz and/or feldspar from this zone. This will take place especially readily if the feldspar and quartz originally in the host rock were utilized in the formation of the pegmatite. The enrichment of hornblende in the external contact areola and the growth of quartz and feldspar in the pegmatite could then be looked upon as complementary processes: feldspar and quartz replace hornblende in the pegmatite, and the liberated hornblende substances replace quartz and feldspar in the external contact zone, the elements of the latter minerals being added to the growing pegmatite body. On the other hand if the original minerals of a rock are not utilized during the evolution of a pegmatite space will be inadequate for fixation of superfluous elements in the region just outside the pegmatite. Consequently no trace of these elements may be found by simple field methods. This mechanism explains why sillimanite may be enriched outside pegmatites in sillimanite schists, and muscovite outside pegmatites in moscovite schists. The petroblastically growing pegmatites which in part or totally replace the sillimanite or muscovite schists, respectively, cannot digest all the aluminum in

these aluminous host rocks. Some of the superfluous alumina may then become fixated in the external contact zone if conditions are favorable.

The contact areolae around the Greenlandic pegmatites may be more complex than this. In micaceous basic inclusions one finds narrow pegmatitic veins or dikes which may be surrounded by zones of coarse hornblende. The pegmatites are plagioclase-quartz bodies with little or no potash feldspar. It is not plain which reactions created these contact zones (Fig. 17).

A typical external contact zone rich in hypersthene is developed around some pegmatites in olivine rocks. In the same pegmatites an internal contact zone of hornblende ± diopside ± biotite has developed around the pegmatite proper which consists of quartz and feldspars (Fig. 11).

Many of the pegmatites studied exhibit structural relations which prove that the pegmatites must have formed by volume-by-volume replacement of their host rock, whereas others have provided their space by dilation. The well-known Goodspeed (1940) criterion for nondilation emplacement of obliquely crosscutting pegmatites is very commonly exhibited by the Greenlandic pegmatites (Pl. 8, fig. 2; Pl. 9, fig. 3). The Goodspeed criterion on nondilation is not conclusive unless the pegmatite cuts two (or more) inclined planar structures in the host rock without offsetting (for example a conformable layer in a banded gneiss and a prepegmatite crosscutting vein). If the pegmatite cuts obliquely only one set of planar structure without offsetting, it is possible that the dilation took place in a direction parallel to the planar structure of the host rock. However, obviously dilated dikes, such as the numerous diabases in the area, generally show offsetting of oblique-banded structures in a fashion which indicates that the general dilational motion of the dike-occupied fissures is approximately perpendicular to the walls.

The Greenlandic pegmatites are often formed in connection with faults and mylonites, so that relative motion of the adjacent rocks parallel to the sides of the pegmatites is apt to make the geometric criteria for replacement difficult to apply.

The structure of host rocks remains as ghostly relics in some transgressive as well as conformable pegmatites (Pl. 7, figs. 1, 2; Pl. 8, fig. 1). In many pegmatites, though, there are no obvious macroscopic traces of host-rock structures. Yet the geometry of the pegmatite and host-rock structure show volume-by-volume replacement.

Pegmatization in Greenland was in part associated with faults, and drag-fault structure is commonly present in rocks adjacent to pegmatite bodies.

The pegmatites are confined bodies without visible avenues for supply from conventional sources (mother magmas). There are reasons for this puzzling fact: (1) mother magmas are not necessary sources; (2) the avenues of supply are represented by every microscopic and submicroscopic path in the host rock along with activated ions, atoms, or molecules can migrate.

Unaltered Host-Rock Fragments in Pegmatites

Inclusions of host rock are common in pegmatites. These inclusions may show parallel linear or planar structures thus possibly indicating that they are remnants of the host rock which has been replaced *in situ* by pegmatites. In several instances, however, host-rock inclusions are rotated in relation to each other and in relation to the adjacent rock. This need not indicate magmatic origin of the pegmatite because replacement of a breccia with rotated fragments can readily result in such structures. Also growth of pegmatitic minerals around rotating inclusions in gneisses can explain the occurrence of rotated rock fragments in nonmagmatic pegmatites. The crucial point is not the state of rotational displacement of inclusions but rather whether or not they originate from the immediately adjacent host rocks, and whether or not they are floating unsupported in the pegmatite body. In Greenlandic pegmatites all rock fragments observed belong to the adjacent host rock. The snowball-made pegmatites which enclose basic inclusions, but yet occur in quartz-feldspar gneisses, seem to be exceptions to this; however, the basic inclusions in these pegmatites originally were inclusions in the gneiss and

thus really belong to the host rock of the pegmatite. These observations show that the pegmatite material was not emplaced by bodily motion of a liquid through channels wider than the cross sections of the inclusions. A bodily injected melt will now and then carry rock inclusions picked up at some distance from the site of the magmatic rock; this is the case with diabase dikes and other magmatic dikes like rhomb porphyry.

The enclosed rock fragments in the studied pegmatites are generally floating unsupported in the pegmatite. This shows that the pegmatites have never been stagnant liquid bodies from which the minerals slowly precipitated, for in such liquid bodies enclosed rock fragments would either rise or sink and thus either cluster on the roof or at the floor of the chambers. (It is very unlikely that the density of the rock fragments would be so close to that of the magma that vertical motion would not occur.)

Internal Texture and Structure of the Pegmatites

The internal structure and texture of the studied pegmatites range within wide limits. Some of the lenticular and equidimensional pegmatites are concentrically zoned, and some of the sheet-shaped dikes exhibit zones parallel to the walls. Many pegmatites do not show zoning except perhaps for a narrow border zone which usually is more fine grained and texturally different from the bulk of the pegmatite. At a few places a coarse-grained border zone has been found (Pl. 9, fig. 1). Some pegmatites consist of a rather uniform aggregate of quartz and feldspars ± biotite, whereas others contain large crystals mixed unevenly with fine-grained mineral aggregates. Euhedral feldspars are found in some pegmatites, but anhedral feldspar and quartz are much more common. Graphic granite intergrowth of quartz and potash feldspar occurs, and a more irregular macro-myrmekitic intergrowth of quartz and acidic plagioclase is seen. Macroscopically obvious replacement phenomena are observed—*e.g.*, replacement of hornblende by a quartz-feldspar aggregate (Pl. 9, fig. 4) or of tourmaline by garnet. The latter forms pseudomorphs after radiating tourmaline needles. Very commonly relic structures of the host rock occur in the studied pegmatite. Conformable pegmatites may contain thin stringers of host rock parallel to the pegmatitic borders, and the host rock may continue into crosscutting pegmatite as ghostly shadows. The mineralogical development of pegmatites crossing through layered complexes is sometimes influenced by the immediately adjacent rock layer. In most pegmatites country-rock structure and/or mineralogy is not reflected to such an extent that a superficial field study reveals the relationships.

Various signs of strain occur in some pegmatites. Thus undulatory quartz is the rule rather than the exception; biotite is typically bent, and tourmaline broken. Schistosity and mylonitization have developed in some of the pegmatites studied.

Most zoned pegmatites have a pure quartz core. Such cores are perhaps the most conspicuous feature of the zoned pegmatites studied. Generally only two or three zones are developed: a fine-grained border zone of quartz and plagioclase, with some potash feldspar and ferromagnesian minerals; a coarser-grained quartz-feldspar intermediate zone with or without various ferromagnesian minerals; a quartz core. The core may contain biotite flakes (Pl. 9, fig. 2) or potash feldspar grains, but pure quartz cores are most typical (Pl. 5, fig. 3). A good example of well-developed zoned pegmatite is a 1.3-m thick crosscutting lenticular body occurring in hypersthene gneisses on a little island west of the inlet to Söndre Isortoq fiord. This pegmatite has a small-grained border zone consisting of dark-gray plagioclase, bluish quartz, a little alkali feldspar, and magnetite; coarser plagioclase, alkali feldspar—in places developed as graphic granite—, and quartz in an intermediate zone; and an almost pure quartz core with a little feldspar. Biotite is scattered throughout the pegmatite and also in the quartz core. This pegmatite is surrounded by a 2–3-m thick contact zone in which the hypersthene in the gneiss is altered to biotite.

The muscovite-rich border zone which appears to be so typical for the pegmatites described from United States (Cameron *et al.*, 1949, p. 24–42) is not present in the Greenland pegmatites, probably because of the higher

temperature of formation. For such variable rocks as the studied pegmatites some pictures of typical bodies are more informative than the most lengthy descriptions (Figs. 12–16).

In some cases it is difficult if not impossible to decide whether the outer zone of some pegmatites belongs to external reaction or extraction areolae or to the internal border zones. This is true not only of pegmatites with diffuse transitional contacts, but also of many exhibiting macroscopically sharp contacts (Fig. 11). The reason for this seems to be that the interaction between pegmatite material and adjacent rock often creates the external and the internal contact zones simultaneously, some of the elements in the internal border zone being taken from the adjacent host rock and some of the elements in the external contact zone being supplied from the pegmatite body. Under such circumstances it is meaningless to distinguish between external and internal contact zones.

FIGURE 11.—PEGMATITE IN DUNITE

Pegmatite in dunite body enclosed in gneiss, Söndre Isortoq. Diagonal hatching: feldspar-quartz-biotite pegmatite; black: hornblende-diopside zone; dotted: hypersthene; white: olivine (with accessories). Length of pegmatite body as shown is about 3 m.

The behavior of mixed crystals through the sequence of zones in the Greenlandic pegmatites has not been studied intensively. Biotite and hornblende are rather common minerals in the studied pegmatites; these minerals show extensive isomorphic exchange of elements. The macroscopic color of these minerals is also rather sensitive to compositional changes, particularily to changes in the Fe/Mg ratio. Differences in color of biotite or hornblende

PLATE 9.—PEGMATITES

1. Quartz-feldspar dike with coarse-grained pegmatitic borders and fine-grained aplitic central zone. Godthåb fiord. Photograph, H. R., 1951

2. Large flakes of biotite in quartz core of pegmatite in amphibolite, small peninsula on north side of outlet of Söndre Isortoq. Photograph, M. L. Ramberg, 1949

3. Crosscutting quartz-rich pegmatite showing nondilation geometry relative to banded host rock, a little south of Söndre Isortoq. Photograph, M. L. Ramberg, 1949

4. Plagioclase-quartz pegmatite containing hornblende with cores replaced by plagioclase and quartz, Netsilik. Photograph, M. L. Ramberg, 1949

PLATE 10.—PEGMATITES

1. Crosscutting pegmatite with diffuse borders in gneiss. Note concentration of biotite flakes in pegmatite in the continuation of hornblende-biotite rich layer in host gneiss. Black circular body on pegmatite is lens cover for camera about 5 cm diameter. Proximity of Christianshåb. Reproduced from Kodachrome, H. R., 1947

2. Boudins of diopside skarn with pegmatites in bytownite carrying banded gneiss, Amitssuarsoralaq, south of Söndre Isortoq. Photograph M. L. Ramberg, 1949 (*see* key sketch, Fig. 15)

3. Zoned quartz-calcite vein in low-grade shale, Lofjorden, Tröndelag, Norway. c = calcite, q = quartz. Photograph, H. R., 1954

4. Pegmatitization of amphibolite boudins and country gneiss in the pressure-shadow regions between the boudins, Ikertoq. Photograph, H. R., 1946

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PEGMATITES

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PEGMATITES

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

PEGMATITES

could not be detected throughout the various zones in the pegmatites studied. Even biotite flakes more than 1 meter wide did not show zoning of color which could indicate compositional zoning.

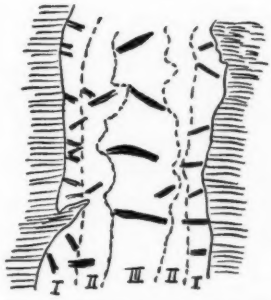

FIGURE 12.—ZONED PEGMATITE

Zoned pegmatite in hornblende-bearing plagioclase gneiss with amphibolite schlieren, Söndre Isortoq. I: dark gray to black plagioclase, quartz, and biotite; II: coarse potash feldspar and quartz; III: quartz core with biotite. The same pegmatite continues into granodioritic gneiss, but there the border zone (I) consists of potash feldspar and quartz.

In the zoned pegmatite (sample 29871) shown in Figure 14*b* a single biotite flake extends across the whole pegmatite. Yet optical check yielded the same index of refraction across the flakes ($\gamma \sim 1.650$). Plagioclase crystals from border and center in the same pegmatite have uniform index of refraction, $\gamma \sim 1.548$, which corresponds to $An_{25\pm}$.

Pegmatite 35622 (Fig. 14*a*) contains a dark-gray antiperthite with yellow "flames." The dark portions have the same index of refraction in the center of the crystal as along the margin, $\gamma \sim 1.548 \pm 0.002$—*i.e.*, $An_{25\pm}$. No difference could be detected between the plagioclases along the border of the pegmatite and those in the central zone. The yellow altered "flames" in the dark-gray plagioclase are somewhat more albitic, $\gamma \sim 1.544 \pm 0.002$. The host rock, however, has a considerably more basic plagioclase, $\gamma \sim 1.557$: An_{35}.

Another pegmatite in amphibolite (sample 29807) from Amitssuarsoralaq south of Söndre

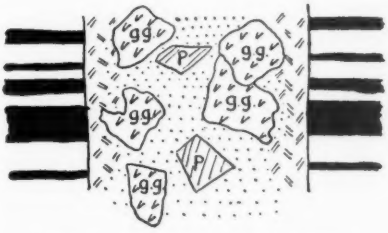

FIGURE 13.—PEGMATITE IN GNEISS

Three-meters-wide pegmatite in banded gneiss, Ameraliq fiord. P: potash feldspar; g.g.: graphic granite; dotted: quartz; short double dashes: small-grained quartz and feldspar.

Isortoq fiord has also been checked for composition of the plagioclase. The plagioclase in the border zone is a little more basic than the plagioclase from the central portions; γ is a little above and slightly below 1.549 respectively. Sörensen (1953, p. 50), however, reports some changes in the composition of pyroxene in pegmatites from Greenland.

The behavior of graphic granite in the pegmatites is significant because its occurrence shows that it does not represent an eutectic mixture. The occurrence in a single pegmatite of large graphic granite crystals with equally large crystals of quartz-free perthite and of pure quartz (Fig. 13) is hard to reconcile with the hypothesis that the graphic granite was the solidified eutecticum. Before the quartz-feldspar eutecticum is reached either feldspar or quartz will form depending upon the original composition of the melt, but both cannot crystallize together. It is not legitimate to assume that quartz will crystallize after the

PLATE 11.—PEGMATITES

1, 3. Hornblende-bearing quartz-feldspar pegmatites in amphibolites. Localities: Fig. 1: proximity of Egedesminde; Fig. 3: innermost part of Arfersiorfik fiord. Photographs, H. R.

2. Feldspar pegmatite with crystals of diopsidic pyroxenes (black crystals in figure) in host rocks of skarn and crystalline limestone, Marble Island, Nagssugtoq. From Kodachrome, H. R., 1947

4. Crosscutting pegmatite with sharp borders. Pegmatite offset by conformable fault. Geometric relation between pegmatite and light-gray band in country rock indicates nondilational emplacement. Godthåb fiord. Photograph, H. R., 1951

quartz-feldspar eutecticum because the eutectic mixture is the last to solidify of any complex melt. Hence it appears that the graphic granite in the pegmatites is not the eutectic mixture of a quartz-feldspar melt.

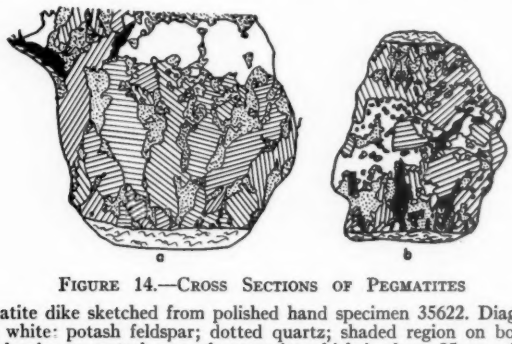

FIGURE 14.—CROSS SECTIONS OF PEGMATITES

a Portion of pegmatite dike sketched from polished hand specimen 35622. Diagonal hatching: plagioclase; black: biotite; white: potash feldspar; dotted quartz; shaded region on bottom is country rock. Section extends from border to central zone of pegmatite which is about 25 cm wide. Fiskefiord district.
b Cross section of pegmatite dikes sketeched from polished section 29871. Diagonal hatching: plagioclase; black: biotite; white: potash feldspar; dotted: quartz; shaded regions on top and bottom are country rock. Thickness of dike about 14 cm. Fiskefjord district.

MINERALOGICAL AND CHEMICAL INTERRELATION BETWEEN PEGMATITES AND HOST ROCK

After having studied thousands of pegmatites scattered throughout the coastal region of West Greenland extending over 5 degrees of latitude the writer has concluded that there is a general mineralogical and compositional interdependence between pegmatite and host rock. This interdependence is very obvious in many cases; in other cases it may be more obscure and difficult to detect by field studies.

Certain mineralogical similarities between pegmatite and host rock prove that the pegmatite and host rock have formed or evolved under nearly the same P-T conditions, so near indeed that differences cannot be detected by means of the stability conditions of mineral assemblages as established by detailed mineral-facies studies. Mineralogical likeness of this kind between pegmatite and host rock occurs for example where sillimanite is found in pegmatites in sillimanite gneisses, where hypersthene occurs in pegmatites in hypersthene gneisses, where granulite-facies garnet (poor in manganese, rich in magnesium) occurs in pegmatites in granulite-facies regions, or where epidote occurs as a major mineral together with plagioclase of similar composition in pegmatite and host rock. The writer has studied from this angle the pegmatites in granulite-facies and in lower-facies regions in West Greenland (Ramberg, 1948, p. 44–54). This study shows that the mineral assemblages of the pegmatites in granulite-facies complexes correspond to the P-T conditions of granulite facies, whereas the minerals of pegmatites in lower-facies complexes correspond to lower facies. The minerals and mineral assemblages employed in the original study were rhombic pyroxene, manganese-poor garnet, and typical granulite-facies hornblende (brown titaniferous and iron rich). In addition the following minerals and mineral characters are useful P-T guides: sillimanite, cordierite, the detailed structure of the feldspars (types of perthite, color, etc.) Antiperthite and mesoperthite—*i.e.* $Or_{50}Ab_{50}$—are typical for granulite-facies conditions, as is a greenish-gray color of the plagioclase and a certain bluish quartz color.

Such relative temperature-pressure indicators show that the greatest number of pegmatites within the three granulite-facies areas studied (the Isortoq, the Kangamiut, and the Kangeq complexes) have developed under the condition of granulite facies which is characterized by high pressure and rather high temperature (ca. 600°C). Only two pegmatites found contain muscovite within these regions and thus possibly (but not necessarily) are formed at lower temperatures than those corresponding in the regional metamorphism of the country rocks. One of the muscovite pegmatites occurs

at the abandoned Isortoq outpost, north of Holsteinsborg, and another a few kilometers north of Kangeq outpost in the Kangeq complex.

In the Isortoq granulite-facies complex there are various kinds of pegmatites, and not all of them carry minerals suited for metamorphic-facies determination. Thus some pegmatites consist of but quartz, feldspar, and biotite, mineral assemblages which can be stable in amphibolite as well as in granulite facies. Only detailed structures of the feldspars (antiperthite with a relatively basic host plagioclase, and mesoperthite indicate granulite facies) and possibly the character of biotite could reveal the P-T conditions under which these pegmatites developed. A few microscopic checks of the feldspar in such uncertain pegmatites revealed the presence of antiperthite and mesoperthite thus indicating evolution at the P-T conditions of granulite facies.

Pegmatite 35622, for example, does not contain ferromagnesian minerals which could definitely place it in granulite or amphibolite facies. It contains, however, cloudy antiperthitic plagioclase and a fine perthitic mesoperthite with about $Ab_{40}Or_{60}$. These minerals therefore place the pegmatite in granulite facies which also is represented by the country rocks of the Kangeq complex.

Samples 388 and 37512 are biotite-garnet-bearing pegmatites in garnet-sillimanite gneiss from Nordre Isortoq. The garnets are poor in manganese but rather rich in magnesium (Ramberg, 1948, p. 45–46), as required of the granulite-facies garnets when potash feldspar and biotite also are present. Both pegmatites have well-developed antiperthites with $\gamma \sim 1.548$.

Sample 37976 is a biotite-bearing pegmatite in the Isortoq granulite-facies complex. The pegmatite has antiperthite plagioclase ($\gamma \sim 1.548$).

The ferromagnesian minerals from pegmatites in the amphibolite-facies regions conform closely to the degree of regional metamorphism in the country rocks. Manganese-rich garnet and lack of hypersthene and of brownish typical granulite-facies hornblende characterize these pegmatites (Ramberg, 1948, p. 44–54). Ellitsgaard-Rasmussen (1954, p. 62–65) arrived at the same conclusion for the pegmatites and feldspar-bearing quartz veins in the low-grade schists and greenstones north of Egedesminde.

Significantly, then, the pegmatite bodies in West Greenland developed at P-T conditions very similar to those characterizing the regional metamorphism in the country rocks. This is an important piece of information to consider when the origin of pegmatites is discussed.

In many cases the chemical composition of the pegmatite is affected by the country rock(s).

(1) Graphite-bearing schists, which are widespread in the Isortoq granulite-facies complex, contain pegmatites with large flakes of graphite (Localities: Utoqait, southeast of Holsteinsborg; Kangerdluarssuq and Isortoq north of Holsteinsborg, and in the fiord Nagssugtoq).

(2) In cordierite-garnet gneisses in Nagssugtoq pegmatites rich in cordierite and garnet are common. Cordierite-anthophyllite schists occur in a small fiord branch in Godthåb fiord. These schists contain small pegmatite veinlets very rich in cordierite.

(3) Sillimanite occurs in some pegmatites in sillimanite-rich granulites in Nordre Isortoq and Nagssugtoq.

(4) Diopside, scapolite, hornblende, and sphene are regular constituents of pegmatites in skarn rocks or in marbles, all over the area (Pl. 14, fig. 2). Very basic plagioclase may also occur in such pegmatites. In the quartz-feldspar pegmatites in a diposide skarn boudinage (Fig. 15; Pl. 10, fig. 2) the plagioclase ranges between bytownite with $\gamma \sim 1.570 \pm 0.003$ and labradorite with $\gamma \sim 1.560$ (sample 29818). The incompetent adjacent gneiss is a gray bytownite-quartz rock with biotite and apatite; the plagioclase shows $\gamma \sim 1.576$ in samples 29816 and 29820 of this gneiss. This gneiss is interpreted as an altered lime-rich sediment because of its association with mica schist and typical diopside skarn layers.

(5) Garnet is especially abundant in almost all rock types in the granulite-facies area around Nordre Kangerdluarssuq and Nordre Isortoq. This is probably related to the peak of regional metamorphism that affected this region. Pegmatites in this region are also unusually rich in garnet.

(6) Some pegmatites which cross contrasted rocks are chemically influenced by the contact

rock. This is true for a pegmatite which crosses granodioritic gneiss and amphibolite near Marrait, north of Christianshåb settlement. The pegmatite is pink and rich in potash feldspar where it penetrates quartz-feldspar gneiss but white and rich in plagioclase where it runs through amphibolite. A pegmatite which crosses through gneiss and amphibolite on the north shore at the mouth of Söndre Isortoq has a quartz-plagioclase (dark-gray plagioclase) border zone (sample 13159, An_{25}) where it penetrates the amphibolite, but a potash-feldspar-quartz border zone (sample 13160) where it penetrates the gneiss.

Hornblende-bearing plagioclase-quartz pegmatites (sample 29885–29886) occur on a little island in the vicinity of Netsilik in the *skaergård* region north of Godthåb. The country rocks are quartz-diorite gneisses with inclusions and bands of amphibolites. Hornblende is commonly enriched in the pegmatites where they cross amphibolite layers or inclusions.

A narrow pegmatite dike on an island south of Atamik in Fiskefjord shows similar features. In this pegmatite biotite is especially concentrated where it crosses through hypersthene amphibolite inclusions in the enclosing hypersthene quartz dioritic gneiss.

In the vicinity of Christianshåb numerous pegmatites occur in the gneisses and schists. Flakes of biotite are concentrated in a pegmatite where it crosses a hornblende-rich band in the gneiss (Pl. 10, fig. 1).

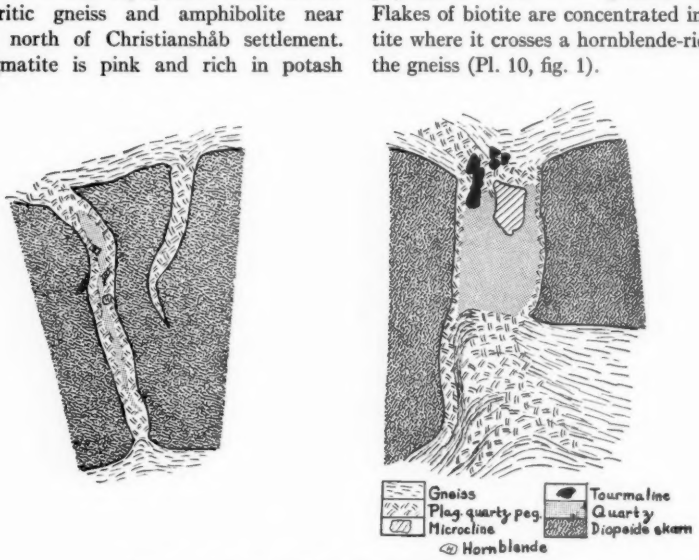

FIGURE 15.—PEGMATITES IN BOUDINAGE STRUCTURE

Pegmatites filling spaces between boudins of diopside skarn in bytownite-bearing gneiss Amitssuarsoralaq, south of Söndre Isortoq.

(7) Most of the Greenlandic pegmatites which occur in amphibolites and other basic rocks or mica schists are white to light gray in contrast to the light-red pegmatites enclosed in the granodioritic genisses. This is especially conspicuous in the Egedesminde district and in the innermost parts of Söndre Strömfjord. The reason for the difference in color is not always investigated, but in checked cases the red pegmatites in the gneisses are rich in potash feldspar whereas the white pegmatites in the amphibolite are quartz-plagioclase rocks. Even the border zone is commonly a quartz-potash feldspar aggregate in the red pegmatites in the granodioritic gneisses.

The white feldspar in the pegmatites in the mica schists may be potash feldspar in some cases for white potash feldspar occurs in these sillimanite-rich schists.

It is especially noteworthy that "large" (veins thicker than a foot or so) hornblende-carrying pegmatites are found only in amphibolites such as for example a large body of hornblende dioritic gneiss in Arfersiorfik, or

the relatively large amphibolite complex on the north side of Alangua and in many of the thousands of smaller amphibolite layers and other basic inclusions in the gneiss (Pl. 11, figs. 1, 3, 4). Smaller pegmatite veinlets, however, may carry hornblende where they occur in

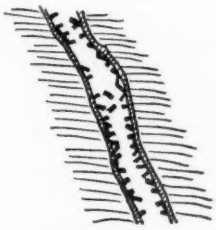

FIGURE 16.—FELDSPAR PEGMATITE IN AMPHIBOLITE

Between Söndre Isortoq and Alangua. Diagonal hatching: hornblende and diopside zone; black: unidentified mineral; white: feldspar. Width of pegmatite about 15 cm.

hornblende-bearing granodioritic gneiss or in gneiss rich in amphibolite inclusions. In the few instances where the hornblendes of pegmatites and enclosing rocks have been studied optically, little or no difference has been detected. Sample 35613 is a pegmatite from the Fiskefjord district. It contains small grains of hornblende, mostly along the border zone. The hornblende has $\gamma \sim 1.672 \pm 0.003$. The host rock is amphibolite with the same kind of hornblende but possibly with a little lower index of refraction, $\gamma \sim 1.670 \pm 0.003$.

Sample 7820 is a little plagioclase pegmatite from a basic inclusion in gneiss in Godthåb fiord. Hornblende, enriched in an external contact zone (Fig. 17), projects into the border zone of the pegmatite and must be considered a mineral belonging to the pegmatite. There is no detectable optical difference between the hornblende in the pegmatite and in the host rock: $\gamma \sim 1.650$ for both.

Another pegmatite in the same district, however, has a hornblende with $\gamma \sim 1.662$, whereas 4 cm away from the pegmatite contact the hornblende in the host rock shows $\gamma \sim 1.654$.

The white pegmatites which abound in the biotite schists in the Christianshåb district carry biotite flakes but no hornblende. The biotites in pegmatite and host schist were checked optically in a few cases. The biotite in pegmatite 37809 has the same kind of biotite as has the host schist, $\gamma \sim 1.634 \pm 0.003$.

Sample 7681 is a banded replacement pegmatite from Godthåb fiord (similar to Pl. 7, figs. 1, 2). The coarse biotite flakes in the

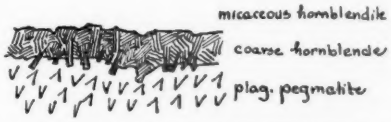

FIGURE 17.—CONTACT OF PEGMATITE

Contact between plagioclase pegmatite vein and biotite-rich hornblendite. Thickness of hornblende zone along contact about 1.5 cm.

central portion of the pegmatite have the same index of refraction ($\gamma = 1.643 \pm 0.003$) as that in the host schist.

In some cases where tourmaline is found in pegmatites in mica schists, this mineral is also a constituent of the schist. Tourmaline is a typical constituent of nonpegmatized schists in Greenland and is supposedly formed from the primary boron in the sediments (Ellitsgard-Rasmussen, 1954, p. 40).

Remarkable cases of chemical dependence of pegmatites and allied veins on host rocks are also well known elsewhere. Thus, for example, quartz veins in aluminum-rich staurolite-graphite schists in Tydal, Norway, contain large crystals of kyanite (samples 35729N and 35720N). Such occurrences show not only that veins are strongly influenced by the chemistry of the host rocks, but also that the quartz vein was not emplaced as concentrated aqueous solutions, for such solutions must be rather rich in alkalies and hence would yield muscovite or paragonite rather than kyanite at the moderate temperature of these schists. However, such chemical interrelations are not always obvious and consistent: we must rather speak of an increased chance of finding certain types of pegmatites in certain types of host rocks. But anybody who discards the observation that pegmatites are influenced by the chemistry of their enclosing rocks as "old wives tales" (Heinrich, 1953, p. 81) rejects in fact a piece of information without which the puzzle of pegmatite formation cannot be solved.

Genesis

Based on observations of the studied pegmatites and known properties of magmatic liquids the writer has rejected the hypothesis that these rock bodies formed by injection and crystallization of a melt. We shall list some observations which make any magmatic theory at best an awkward and inconsistent explanation of the formation of the studied pegmatites.

(1) The pegmatites are generally confined bodies without visible connections with magmatic sources, a fact especially emphasized in a study of Norwegian pegmatites by Andersen (1931) and later by Ramberg (1943, M. A. thesis, Univ. Oslo).

(2) No unquestionable magmatic mother rocks and very few massive, homogeneous granitic bodies have been found in the area.

(3) Unquestionable evidence of pegmatite emplacement by volume-by-volume replacement of host rock is found in many places.

(4) Emplacement by a very slow growth and expansion of the crystalline pegmatite bodies has been inferred from structural observations. This is inconsistent with the thesis that the space now occupied by pegmatite was originally filled by a magma which solidified into the pegmatitic rock.

(5) The general zoning of pegmatites is not consistent with the empirically known, or the theoretically expected properties of silicate magmas. Especially the commonly occurring pure quartz core is contrary to the expected late-crystallizing fraction of any complex melt and of natural silicate melts in particular. Instead of being purely monomineralic the last portion to solidify should be composed of the eutectic polymineralic assemblage of the multicomponent melt. The fact that mixed crystals such as plagioclase, hornblende, and biotite do not appear to change consistently from margin to center in the pegmatites is also contrary to the expected result of fractional crystallization of a melt. In recent years pegmatites have been studied in great detail by a number of students in United States (Cameron *et al.*, 1949; 1954; Page, Stoll, Hanley, Heinrich, and many others). However, up to now no satisfactory explanation of the origin of the American pegmatites has been proposed. For example, Cameron *et al.* (1954, p. 31) found very slight difference in composition of plagioclases in various zones in pegmatites in New England. Generally the plagioclase in the intermediate zone was only a couple of per cent less basic than the border-zone plagioclase; in some instances there was no difference, in others it was about 5 per cent. Such inconsistent and small change in the plagioclase composition from margin to center does not support the hypothesis of fractional crystallization. The zoning in unquestionable magmatic dikes is negligible or lacking, whereas many kinds of veins, secretions, concretions, and other congregations in low-grade schists are distinctly zoned. *In general metamorphic differentiation rather than magmatic differentiation seems more capable of creating zoning of the segregated rock bodies.*

(6) The coincidence between temperature of crystallization of pegmatites and grade of regional metamorphism in the country schist-gneiss complexes is not readily explainable if the pegmatites were injected magmatic bodies, because such bodies would have a solidification temperature range determined by their bulk composition. Magmatic dikes crystallize at temperatures independent of the degree of the regional metamorphism that affected the host rock.

(7) The existence of all mineralogical gradations between quartz-feldspar pegmatites and metamorphic-metasomatic quartz veins or calcite-quartz veins, and the great similarity in structure and texture of these veins, clearly points toward some genetic processes other than magmatic differentiation.

(8) The pronounced mineralogical-chemical influence of pegmatites by host rocks is perhaps not inconsistent with a magmatic theory, but such relationships between dikes and host rocks are not typical of unquestionable magmatic dikes.

(9) The secondary foliation and crumpling impressed on adjacent rocks by many pegmatites show that the pegmatites were mechanically stronger (more competent) than the host rocks. Since the latter were crystalline bodies the pegmatites also must have been crystalline under their evolution. The same can be inferred from ptygmatic folding and

pinch-and-swell structure. In the magmatic theory this common structural feature around pegmatites must be explained by some secondary processes subsequent to solidification; in the metamorphic-metasomatic theory such structures follow naturally, because in this theory the pegmatites were solid all during their growth.

(10) Pegmatites occur typically in regional metamorphic schist-gneiss terrains or in plutonic rocks of uncertain origin. (*See* Andersen, 1931; Cameron *et al.*, 1949, p. 4–8). Quartz-feldspar pegmatites are not found in low-grade or unmetamorphosed sediments although diabase dikes and rhomb porphyry dikes are common in such rocks. (Some pegmatites occur in unquestionable magmatic rocks such as, for example, the nepheline syenite pegmatites of the Oslo region in Norway; Brögger, 1898, p. 1–100. Such pegmatites have no doubt gained their constituents from the enclosing magmatic rocks but the detailed mechanism of formation may be similar to that which creates pegmatites during regional (or plutonic) metamorphism. After the magmatic massifs of the Oslo region solidified, joints and fissures opened, and the mobile constituents of the still-hot but crystalline complex diffused toward the low-pressure sites and became fixated there.[1] Thus pegmatites of magmatic parentage can be explained by the same mechanism employed in the metamorphic-metasomatic theory of pegmatites.)

(11) Few pegmatites occur in the granulite-facies areas in Greenland; most bodies are found in amphibolite to epidote amphibolite-facies regions. This means that the majority of pegmatites formed below 600° and consequently below magmatic temperature.

To many students of pegmatites the above points may mean nothing because it is taken for granted that the pegmatites are magmatic, and a number of mutually incompatible properties are ascribed to the pegmatite magma in order to explain some of the observed facts. On some occasions this remarkable liquid must be highly fluid like liquid helium in order to penetrate invisible avenues in the host rocks; at other times it suddenly becomes extremely viscose in order to bulge out as conformable thick lenses in schists or crumble the country crystalline rocks.

On some occasions the pegmatite magma is claimed to contain great quantities of water to explain the growth of huge crystals, but in other instances the same magma must be very dry indeed because anhydrous minerals like hypersthene, sillimanite, kyanite, and pyrope-garnet form instead of hornblende and micas, yet such "dry" pegmatites are not conspicuously less coarse-grained than "wet" pegmatites. Another remarkable property of the hypothetic pegmatite magma is its ability to change its solidification temperature range, not depending upon the composition of the magma itself but controlled by the regional metamorphism of the country-rock complexes. Are we not just requiring too much of a magma; should we not hunt for explanations of pegmatite formation which are more self-consistent and contain fewer contradictory assumptions? The writer proposes a simple and consistent theory to explain the origin of the West-Greenlandic pegmatites (and many other pegmatites as well). This is a metamorphic-metasomatic theory in which diffusional transfer is essential. Some advantages of this theory are: (1) It needs no visible avenues toward the pegmatitic bodies. (2) It needs no magmatic mother massif. (3) It explains the coincidence in P-T between host rock and pegmatite. (4) It harmonizes with the observations that most pegmatites are chemically and mineralogically influenced by their country rocks. (5) It does not require that a pegmatite body derived from a smaller or large volume of the country rocks shall have the same composition as the nearest host rock. Unlike mobility of the various rock constituents gives rise to a pronounced compositional differentiation during the evolution of the pegmatite. (6) It is not inconsistent with the special kind of zoning in pegmatites, for instance the pure quartz core. In low-grade slates and calcareous sandstones secretion veins identical to pegmatites in texture, but with carbonate-rich border zones and pure quartz cores, shown in Figure 3 of

[1] The writer has discussed this point with Oftedahl who has studied the pegmatites of the Oslo region. It appears that Mr. Oftedahl's observations and inferences are consistent with a mechanism of formation closely similar to that described in this paper.

Plate 10, are very common. This shows that metasomatic veins not related to magmatic melts develop pure quartz cores. (7) It explains why pegmatites consolidate in joints and shear zones, why any potential openings in deep-seated rocks become immediately closed by quartz-feldspar minerals or other mobile minerals, and why pegmatites often develop pinch-and-swell structure or ptygmatic folding. (8) It gives a reason for the great similarity in shape and structural features between pegmatite bodies and all kinds of metamorphic vein-shaped low-temperature segregations such as quartz veins, carbonate veins, epidote veins, and alpine veins. (9) It explains why pegmatites are confined to regional metamorphic terrains or to plutons.

The theory is based on but two assumptions: (1) that diffusion may occur in rocks during regional metamorphism, and (2) that, under regional metamorphism, the elements constituting quartz and alkali feldspar are generally considerably more diffusive than the constituents of cafemic minerals. With these assumptions the evolution of quartz veins and pegmatites follows logically, as does, for example, the history of the pegmatites in the space between boudins as shown in Figure 15 and Figure 3 of Plate 6. In the metamorphic-metasomatic diffusion theory the evolution runs somewhat as follows: Tension joints are formed in the competent amphibolite and diopside skarn layers because of plastic one- or two-dimensional expansion of the adjacent schists (or quartz-feldspar gneiss) resulting from the compressional stress. The ruptured pieces of the cafemic rocks are affected by tensional drag forces along the contact with the plastically yielding schist, for which reason the pieces of the cafemic rocks will tend to separate. In the fissures which open between the amphibolite boudins two things will happen either contemporaneously or separately: (1) the adjacent incompetent schist or gneiss creeps into the fissure; (2) and/or minerals precipitate in the fissure. In most cases in the studied area both things happen, but one of the two possibilities may predominate overwhelmingly. In Greenland the fixation of minerals in the dilating joints most commonly predominates over the plastic creep of the incompetent rock into the joints. The latter process is overwhelmingly important in very few cases. The mineral consolidates in the joint because the constituents are transported by diffusion. If the vein material is transported by this mechanism we can conclude on theoretical grounds (Ramberg, 1952, p. 92, 216) that any local low-pressure region not only directs the diffusion currents toward itself, but necessarily also functions as a site of consolidation of the diffusing particles. This is independent of whether the diffusion goes on in a stagnant pore liquid, along mineral surfaces and interfaces, or inside their structures. The tendency of fixation at sites of pressure minima is also independent of whether the diffusion affects individual ions and atoms, or larger complex ions or molecules. It is unfortunate for the fluid-flow theory that the low-pressure regions, such as fissures or pressure shadows in lee of rigid inclusions or large mineral grains, are not necessary or even probable sites of precipitation if the transport is in the form of a flowing liquid solution or magma. This type of transport generally needs a drop in temperature in order to precipitate minerals. As temperature and pressure vary independently there is no reason to assume that low-pressure regions also are low-temperature regions, and it follows that the chances for fixation of minerals in fractures and pressure shadows are small if the transporting agency is a bodily flowing liquid. It is in part for this reason that the writer emphasizes diffusion as the essential means of vein formation in metamorphic terrains.

Thus the fissure between the amphibolite boudins is a site to which the most mobile constituents of the surrounding rock complexes are forced to diffuse and to consolidate to mineral species.

This process is self-regulating; the minerals are transported to the fissure and fixiated there only as long as the pressure is lower there than elsewhere. (Some exceptions to this exist.) The diffusional transfer tends to close all kinds of voids in regional metamorphic complexes. Here the theory of diffusional transfer agrees with the observation that unfilled openings are practically nonexistent in metamorphic terrain.

As to the zoning of pegmatites, in particular the quartz core, the following explanation could be offered: Suppose that silica and alkalies are the most mobile substances in

metamorphic rocks, and that alumina is mobile enough to explain the growth of alkali feldspars in pegmatite veins and feltspathization of schists. Calcium, iron, and magnesium are supposedly less mobile. These relative mobilities can be inferred from various behaviors of rocks and minerals during regional metamorphism. The susceptibility to plastic flowage in gneiss-schist complexes increases along the sequence-ultrabasic rocks (olivine and pyroxene aggregates; diopside skarn), amphibolites (depending somewhat upon the mineral composition of these), quartz-feldspar gneisses. Basic anorthosites (bytownites) are much less plastic than quartz-alkali feldspar gneisses. As plasticity depends upon ability to recrystallize, plasticity and chemical mobility must be closely related.

As soon as the joints open a little, silica and the constituents of alkali feldspar consolidate in the joints, thus giving rise to a quartz-feldspar zone along the walls of the joint. Upon further dilation of the joint the fixation of minerals continues but most probably in a displaced ratio. Silica is abundantly available in most metamorphic areas, and as this constituent also is most mobile its transfer to the fissure will not decline much; however, the feldspar constituents (alkalies and calc)—particularly anorthite—are less abundant than silica in most rocks, and it is possible that the early-formed border zone of feldspar along the joint has exhausted the source of feldspar so that the introduction and growth of quartz will predominate in the further evolution of the pegmatite.

In addition to the relative mobility of mineral-forming matter in diffusion as a clue to zoning, it is also significant that the evolution of a pegmatite takes place over very long intervals of time, and consequently that the P-T conditions may vary considerably during the growth.

The mechanism of formation of some pegmatites which have grown by volume-by-volume replacement is related to the formation of secretion pegmatites; many replacement pegmatites have formed in low-pressure regions. Replacement of schist or gneiss by pegmatite is pronounced in some secretion pegmatites formed in boudin interspaces (Pl. 5, fig. 7). It is natural that regions of low pressure, such as must exist in the portion of incompetent rocks which creep into boudin interspaces, or in lee of rigid inclusions in schists and incompetent gneisses, are places of maximum chance for replacement.

The physicochemical reason why quartz-feldspar minerals replace cafemic minerals of amphibolites, gneisses, and mica schists is not always clear but this does not invalidate the field observations that such processes have taken place. Replacement pegmatites must be considered an important detail of the large-scale architecture of the orogenic regions in Greenland. There is evidence for a regional introduction of alkali feldspars with or without quartz into the amphibolite-facies areas, and a corresponding depletion of these constituents from the subjacent granulite-facies regions (Ramberg, 1948; regional granitization and associated compensating degranitization). Many of the rocks in the amphibolite-facies regions are therefore on the point of being "saturated" with respect to alkali feldspar and quartz even if not carrying these constituents as minerals. This is to say: the activities (free energies) of K, Na, Al, and Si (O) in the rocks are sufficiently high to allow alkali feldspar and quartz to form if the nucleation is made possible. At certain sites in these complexes, such as local low-pressure regions, nucleation is more apt to occur than in other regions, and pegmatites can start growing in these places. Material for further petroblastic growth of the replacement pegmatites is furnished by diffusion through the solid rocks as well as along visible and invisible fissures. Clearly also the most mobile constituents of the enclosing rocks will contribute to the evolution of the pegmatites.

The formation of a replacement pegmatite requires that pre-existing mineral constituents not utilized in the pegmatite must disintegrate and migrate away. This is not a readily understandable phenomenon. One possible mechanism arises if the growing minerals utilize some but not all constituents of pre-existing minerals. As an example, assume that biotite in a pegmatite grows at the expense of pre-existing hornblende in accordance with the following reaction:

$$3Ca_2(Mg,\ Fe)_4\ Al_2\ Si_7\ O_{22}\ (OH)_2 + 2K_2O + H_2O = 4K(Mg,\ Fe)_3\ Al\ Si_3O_{10}\ (OH)_2 + 6\ CaO + 9SiO_2 + Al_2O_3$$

where potassium and water are introduced and calcium rendered superfluous, whereas silica and alumina may be utilized in the pegmatite as quartz and various aluminosilicates. The reaction between introduced potassium and water and pre-existing hornblende will extract, as it were, Mg, Fe, and partly Al and Si out of the hornblende structure and incorporate these constituents in the newly forming biotite nuclei. This results in a slight enrichment of calcium in the hornblende close to the growing mica grain, and a consequent rise of the partial free energy of calcium in this region. This forces calcium to migrate out of hornblende and into the adjacent rock if it is not utilized in some other pegmatite minerals. Whether or not calcium will be able to concentrate in the external contact areola of the pegmatite depends upon many intricate conditions as mentioned previously.

If the introduced substances are supersaturated (*i.e.*, if the free energy of the migrating potash feldspar substance, for example, is higher than that of the crystalline potash feldspar under the considered P-T conditions) then the growing minerals can exert a pressure of growth which then theoretically should cause disintegration and chemical removal of enclosing pre-existing minerals. Such a pressure of growth has been called upon to explain the evolution of conformable pegmatite lenses in schists and gneisses (Concretion pegmatites, Ramberg, 1952). The present study of the pinch-and-swell structure and ptygmatic folds, however, suggest that much of the conformity between pegmatites and adjacent incompetent schist or gneiss, or the crumpling around transgressive and conformable pegmatites, is secondary and caused by the fact that the coarse-grained pegmatites, which are not likely to recrystallize, are competent relative to mica schists and many quartz-feldspar gneisses.

That the Greenlandic pegmatites are more abundant in biotite schists than in quartz-feldspar gneisses can be explained by the metamorphic-metasomatic theory. When the constituents of potash feldspar and acidic plagioclase migrate through a granodioritic gneiss they become fixated homogeneously throughout the rock because nucleation is not necessary inasmuch as feldspars are already present throughout the rock. Only where fissures or joints occur does the migrating quartzo-feldspatic material consolidate to pegmatitic veins. In a mica schist, on the contrary, potash feldspar is rare, and nucleation is necessary before the mobile potash constituents can consolidate. Because nucleation is not probable in the bulk of the schist (unless the migrating substances are supersaturated to a considerable degree), potash feldspar and other minerals not present in the schist do not commonly start growing except as pegmatities at certain favorable localities for nucleation—*e.g.*, joints, shear zones, pressure shadows.

The chemical relationship between pegmatite and host rock is affected by migration along the pegmatite vein, particularly if it follows a joint or fault. In such cases the pegmatite should be less "colored" by the host rock, and chemical incompatibility with consequent interactions between pegmatite and host rock would be possible.

Chemical incompatibility between pegmatite and enclosing rock is not inconsistent with the theory of introduction by diffusion through host rock, for the pegmatites often start to grow at slightly lower pressure than that in the bulk of the host rock. Under such conditions the minerals of the pegmatite and host rock may become incompatible when pressure on pegmatite and host rock is equalized.

In the metasomatic-metamorphic pegmatite theory part of the source for the pegmatite material is the rocks in the vicinity, the contribution of each rock depending upon the mobility of its constituents, its distance from the pegmatite, its size and spatial orientation with reference to favorable avenues of transport to the pegmatite, and the stress and temperature conditions under which it exists. The pegmatites supposedly also gain substance from more-distant sources, inasmuch as the mobile matter which is responsible for regional "granitization" becomes fixated to pegmatites in favorable sites.

Many details of the metamorphic-metasomatic theory of pegmatites remain to be studied, but the theory is simpler, more self-consistent, and considerably more capable of explaining the natural occurrence of pegmatites in all their versatility than is the orthodox magmatic theory.

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GEOLOGY OF ISABELA (ALBEMARLE) ISLAND, ARCHIPIELAGO DE COLON (GALAPAGOS)

BY A. F. BANFIELD, CHAS. H. BEHRE, JR., AND DAVID ST. CLAIR

ABSTRACT

Isabela, westernmost and largest island of the Galapagos, lies 650 miles west of Guayaquil, Ecuador; it is shaped like the letter "J", and measures 40 miles at its widest, 70 miles at its longest. It is entirely volcanic; several of its 5 craters and shield volcanoes, about 2½–6½ miles in diameter, have erupted during this century. The lavas, scoria, and tuff are mainly basaltic, except at one coastal locality east of Volcan Alcedo where there are possible andesites. Lava chiefly constitutes the lower, relatively barren slopes of the island; the upper, grassy or lightly wooded parts are more scoriaceous with some pumice, but late flows abound, especially in and near the craters. Shell-rich sands appear on protected beaches but coral reefs are absent.

Outstanding volcanic features include spatter cones, slickensided flows, rifts, crevasses, solfataras, fumaroles, and strikingly linear eruptions. Breaching of cones follows local patterns—e.g., toward the south on the southeasterly side of Isabela. Sulphur deposits are associated with breaching and with vents emitting steam, hydrogen sulphide, and sulphur dioxide. Apparently oxidation of the hydrogen sulphide yields steam and sulphur.

The general relations of the volcanoes and the relative newness of the western ones suggest that the Archipelago is being extended westward from Cocos Island and the mainland of Costa Rica along a southwestward rift line, one of the two predominating crater alinements. Available soundings suggest linkage of the Galapagos group to Central America rather than to Ecuador.

CONTENTS

TEXT

	Page
Field work and acknowledgments	215
Location and climate	216
General geology	217
The five main craters	219
Lesser craters and cones	221
Fissures and rifts	222
Causes of crater breaching	224
Description of flows	225
Recent eruptive history	225
Volcanic sequence and age	226
Petrography	227
Megatectonic features	227
Major lines of vulcanicity	227
Continental connections	230
Solfataras	231
Location and types	231
Description	231
Rock alteration near sulphur vents	233
Origin of the sulphur	233
References cited	234

ILLUSTRATIONS

Figure	Page
1. Relation of Galapagos Archipelago to continental margin	217
2. Isabela Island showing the five great volcanoes	218
3. Diagrammatic map of Isabela Island	224
4. Map of Galapagos Archipelago	229

Plate	Following page
1. Aerial photographs	232
2. Aerial photographs of Sierra Negra crater	
3. Aerial photographs	
4. Aerial photograph of Volcan Wolf	

TABLES

Table	Page
1. Approximate dimensions of craters	219
2. Chemical analyses of rocks, Isabela Island	228

FIELD WORK AND ACKNOWLEDGMENTS

During January–March, 1953 a reconnaissance survey of the mineral resources of Isabela Island was made which involved investigation of the five major craters and subsidiary volcanic phenomena including sulphur occurrences. The field party consisted of A. F. Banfield (leader), Chas. H. Behre, Jr., and David St. Clair, all of the United States, and Captain Guillermo Bixby of Ecuador. The expedition was organized in Guayaquil, the main seaport of Ecuador.

Because of the isolation of the island and the reputedly arid climate, the expedition was of necessity entirely self-sufficient for a period of at least 2½ months. Planning and preparation were difficult because information gave conflicting data regarding conditions to be encountered.

Field work was carried out during the rainy season in 1953—a year of abnormally great rainfall; instead of the anticipated water shortage, rainfall was so heavy as to reduce visibility considerably and make photography and travel difficult. Though lower elevations were dry, in the uplands the cloud base stayed at about 2000 feet above the sea. Local residents stated that such rain had not fallen for many years and that there had been none during the previous year. Similar conditions, generally traced to change in oceanic circulation, have been described by Murphy (1926) and others.

Special thanks are due the officials of the Great Lakes Carbon Company of New York for generously permitting publication. Study of the islands was authorized by the Ministry of National Defense, Quito, Ecuador. Kind assistance was given by Mr. S. H. Dolbear and Mr. Parke A. Hodges of Behre Dolbear and Company in critically reading this manuscript. Professor Arie Poldervaart of Columbia University was most helpful in discussing the petrographic descriptions. Aerial photographs of much of Isabela Island were made available for study and several were released for publication by the Ecuadorian government through the American Air Force.

Location and Climate

The Galapagos Islands (Fig. 1) lie on the Equator about 650 miles west of Ecuador. The group is officially known in Ecuador as the Archipielago de Colon, but is more popularly referred to as the Galapagos Islands, after the large land tortoises characteristic of the larger islands. The archipelago has been famous, especially among botanists and zoologists, ever since excellent descriptions by Darwin (1839; 1891) and later by Wolf (1892; 1895). Its general geography was recently well but briefly summarized (McBride, 1918). During World War II the islands assumed strategic importance and the United States maintained armed forces there for several years, with permission from the government of Ecuador.

The administrative center of the Galapagos Islands is at Puerto Chico (Boquerizo Moreno) a village of about 500 persons, just north of the westernmost point of San Cristobal Island, about 100 miles east of Isabela Island. Here a government wireless station is maintained. All vessels visiting the Galapagos Islands must report to the authorities at Puerto Chico on arrival and departure. Although there are a few small general stores, supplies at Puerto Chico are very limited; here too fresh water was available.

The largest island, Isabela, also known as Albemarle (Fig. 2), measures about 110 miles in length along the central ridge and from 10 to 30 miles in width. The temperature is generally about 80°F during the day; in the mountains at night it may be as low as 50°F. The five volcanoes which form the higher parts of the island are largely cloud-covered. Even in the "dry" season (May–December) their upper reaches receive frequent fine, misty rains.

The only regular means of communication between Isabela and the mainland of South America is a boat, sent out approximately every 40 days by the Ecuadorian Navy, which calls at all larger islands in the group. Coastal freighters and fishing boats occasionally visit Isabela.

The population of Isabela is said to be about 375, made up of some 250 civilian residents and 125 convicts and officials connected with the prison camps of the island. Villamil is the main center of habitation, a collection of some 25 small buildings on the southern coast of the island with a government radio station. The quantity of supplies at this village is small. The national Porvenir prison camp is near by. About 10 miles north of Villamil on the sole well-marked route of transportation, a narrow horse trail in part over rough lava, are the five buildings of Santo Tomas, a small agricultural community and penal colony. A third penal colony is at Alemania, about 30 miles northwest of Villamil. These, with Villamil, represent the only residences or settlements.

The residents live mostly on fish, and locally grow fruits and vegetables and send to Guayaquil coffee which grows wild. At Villamil and

Santo Tomas fresh water is in limited supply from shallow wells.

The lowest slopes are characterized by a vegetation of the desert type, predominantly cactus. At altitudes of 500 feet this is replaced by thorn bushes. Still higher (1500 feet or so) the growth becomes lush and suggests a tropical rain forest, tree ferns growing in sizeable patches at 3500 feet on Volcan Alcedo and banana plantations at 1300 feet near Santo Tomas. The highest levels, where rugged, bear a moderate forest growth but where flatter are characterized by great savannas in which graze herds of cattle and smaller ones of horses, asses, pigs, and goats, preyed upon by dogs—wild descendants of animals once domestic. The differences in flora reflect contrasts in precipitation. Each zone of precipitation and vegetation seems to rise northward. Vertical contrasts in rainfall cause differing degrees of bedrock decay, whether the rock is fragmental material or lava; thus extensive areas of rich soil are developed above 1500 feet, locally sealing the otherwise porous bedrock so that rainwater accumulates temporarily.

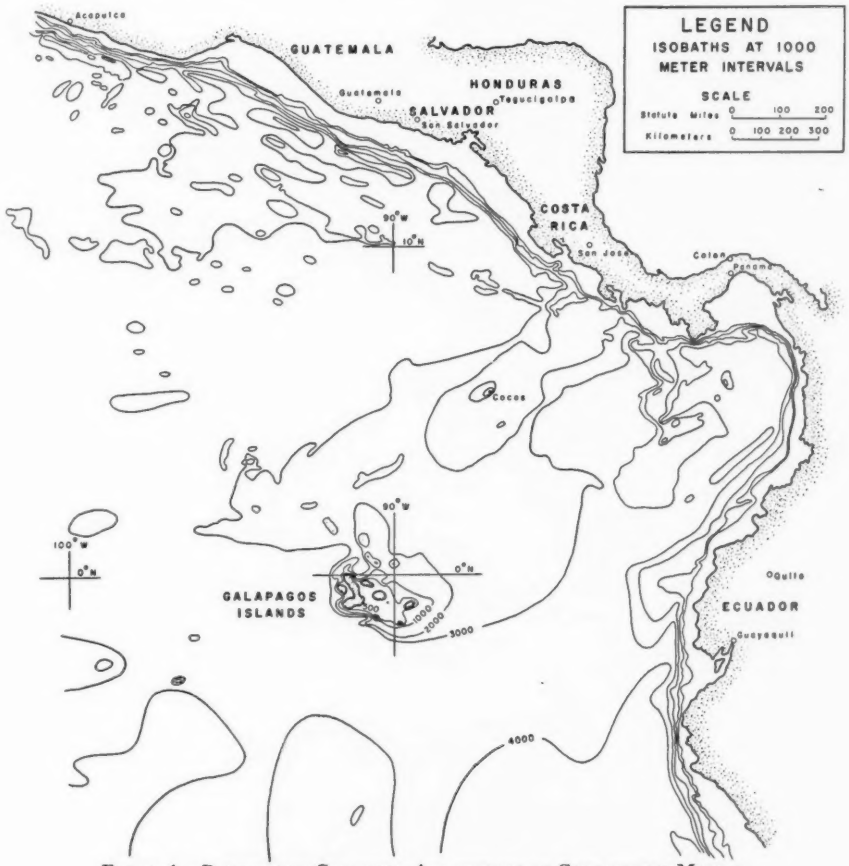

FIGURE 1.—RELATION OF GALAPAGOS ARCHIPELAGO TO CONTINENTAL MARGIN
Isobaths are at 1000-meter intervals

General Geology

Aspects of the geology of Isabela Island have been described by a few visitors, but the only

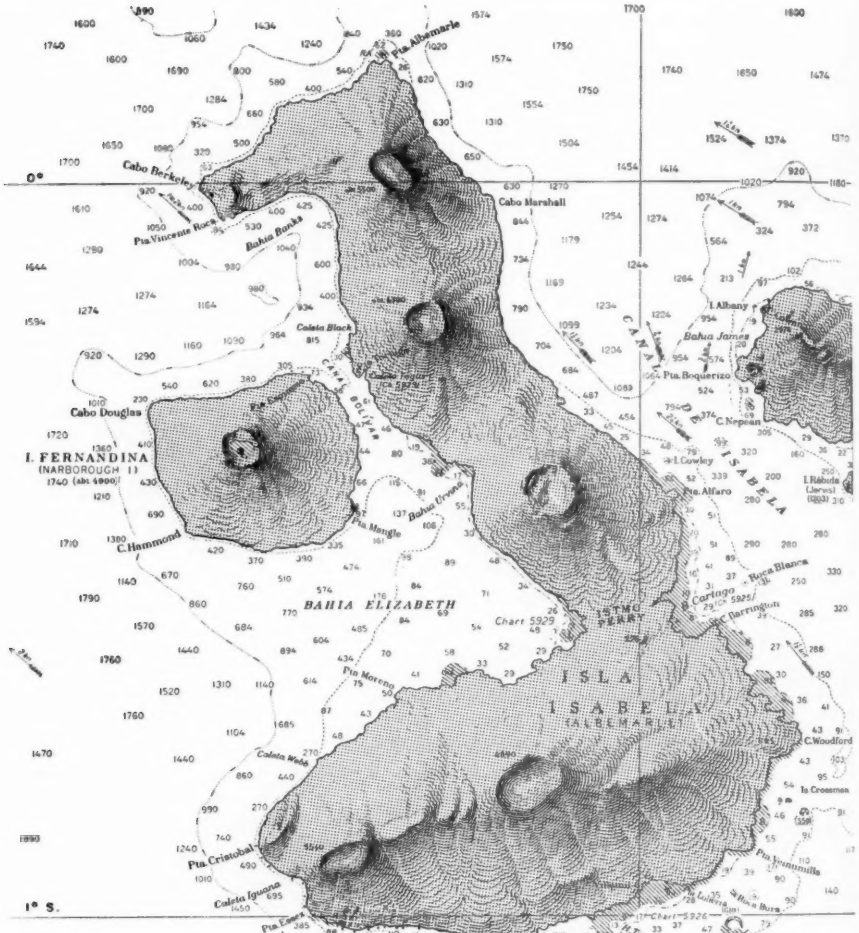

FIGURE 2.—MAP OF ISABELA ISLAND SHOWING THE FIVE GREAT VOLCANOES
Scale 1:005,000 (approx) at equator; from U. S. Hydrographic Office Chart No. 1798

fairly detailed treatments besides those by Darwin (1839; 1890) and Wolf (1895) are the excellent general description by Chubb (1933) and the discussion of the petrology by Richardson (1933).

Isabela Island is composed mainly of ejecta from five great craters so large they could well be designated calderas. From these craters lava flows extend to the sea and toward each other, so that the cones coalesce into one landmass

whose low-lying fingers or steeper cliffs form the coast line. On this base the products—partly flows, partly pyroclastic deposits—of many accessory craters have piled up and form lesser prominences on the slopes of the great shield volcanoes.

In a few places, such as the west shore immediately north and south of Tagus Cove, there are beaches of black sand derived from the lavas. On the southern coast between 2 and 3 miles of beach are made up of calcareous marine fragments. However, except for volcanic clastic ejecta still resting where they fell, sediments make only a very thin and local veneer in a few sheltered parts of the complexly indented shores or form very thin layers between volcanic rocks. The striking absence of coral reefs reflects the prevailingly cooler temperature produced by the Humboldt current, elsewhere flowing generally north but almost westward through this archipelago. Wolf (1895) correctly reported the regular ocean temperature as ranging from 21° to 23°C, about 6° below that near the continental coast of Peru and Ecuador and thus unfavorable for reef-building organisms.

The Five Main Craters

Cerro Azul and Sierra Negra (Volcan Grand) dominate the south lobe of the "J" suggested by the form of the island. The long northward stretch of the island is formed by three other volcanoes. The southernmost volcano has had no widely accepted name; Beebe (1926) called the two to the north Mounts Williams and Whiton, but these names have not been accepted by the Ecuadorian government and are the names of persons living at the time when the names were applied. For both these reasons their use is contrary to the standard procedure in geographic nomenclature. Search was therefore made by the U. S. Board of Geographic Names of the Department of Interior, whose spokesman, Dr. H. F. Burrill (1953, personal communication) reports that official Ecuadorian maps use the following names for the five great craters, the synonyms being given in parentheses:

Name	Approximate Position
Cerro Azul	lat. 0°55′S., long. 91° 26′W.
Sierra Negra (Volcan Grande)	lat. 0°50′S., long. 91° 10′W.
Volcan Alcedo (Volcan Calderon)	lat. 0°25′S., long. 91° 09′W.
Volcan Darwin (Mt. Williams)	lat. 0°12′S., long. 91° 20′W.
Volcan Wolf (Mt. Whiton)	lat. 0°01′N., long. 91° 23′W.

Table 1.—Approximate Dimensions of Craters

	Diameter average, miles	Altitude American Air Force Map (No. 9538)	Altitude U. S. Navy Hydrographic Office Chart (No. 1798 and related series)
Cerro Azul	2¾	5540	5540
Sierra Negra	6	4890	4890
Volcan Alcedo	6¼	3860 (?)	3650
Volcan Darwin	4½	4600 (?)	4300
Volcan Wolf	4½	6100 (?)	5600

A sixth crater, separated from Isabela Island by Canal Bolivar (*see* Fig. 2), forms Isla Fernandina (Narborough Island). It was not visited by the expedition but appears similar to the major volcanoes of Isabela.

The smallest crater on Isabela, Cerro Azul, is oval in outline, measuring about 2¾ by 1 mile; the others are all more nearly circular. The highest points of the crater rims above the sea as reported are given in Table 1, all figures being approximate.

The angle of slope of these volcanoes is, on the average, gentle, but varies considerably among the different volcanoes and in the several parts of each. The average on Sierra Negra and Volcan Alcedo is less than 5° and on the lower flanks generally does not exceed 2°. On the other hand, Cerro Azul has a slope of nearly 10° in its upper portion, and Volcan Wolf slopes 35° in the upper 2000 feet. The entire eastern slope of Volcan Wolf, in fact, has an average angle of 25°, rising to over 5000 feet in a horizontal distance of about 1.8 miles, the upper parts being the steepest.

The depths and slopes inside the crater rims likewise vary. The crater of Sierra Negra has a depth of approximately 350 feet, but those of Volcan Alcedo and Volcan Darwin approach 1000 feet and that of Volcan Wolf, 1500 feet.

The inner walls are generally steep, ranging from 15° to 37° at Volcan Alcedo, and are almost vertical at Volcan Wolf, Volcan Darwin, and Cerro Azul. Benches are not uncommon on the crater walls, in part, apparently, the result of landslides or subsidence of the innermost parts of the craters, as suggested on the northwest wall of Sierra Negra. A similar but far more extensive bench rims almost the entire innermost crater of Volcan Alcedo at an elevation some 400 feet above the crater floor, and the southern part of Volcan Darwin has a bench about 200 feet above the crater floor. Volcan Wolf has such a bench 400 feet or so above its floor on the west side.

Examination of the true-scale profiles of Cerro Azul, Sierra Negra, and Volcan Alcedo confirms their "shield" form, resulting largely from extrusion of fluid lavas. By exaggerating the vertical scale the first two volcanoes are made to show a fairly abrupt change of slope part way up the flank. At Cerro Azul this change is at 2000 feet; above that elevation outcrops that are clearly lava virtually disappear. The slope extends upward toward the crest in the last mile or less of ascent at angles of 5° or less, and in this stretch only dark cinder, scoria, and tuff are found, except for small patches of recent lava. Admittedly the scoria may represent merely the top of a flow. On Sierra Negra, conditions are similar, lava disappearing at an altitude of 500 feet. On Volcan Wolf, lavas make up the only rock exposures to an altitude of about 2000 feet; at higher altitudes, the flows give way progressively to pyroclastic deposits, concomitantly with a decline in angle of slope to an average of less than 10°; at 2700 feet the slope angle increases abruptly to 33° and above this the cone has the steeper slope and form so commonly seen in other volcanic regions where fragmental materials are the chief constituents of cones. In contrast, Volcan Alcedo shows no transition from a lower lava zone to an upper pyroclastic zone. Similar relations are found on Volcan Darwin, which has an almost constant slope of 18°.

Changes in vegetation and soil cover coinciding with changes in slope have been ascribed by Wolf solely to climatic contrasts at differing elevations, resulting in soil differences which masked the underlying, essentially uniform lavas. This explanation, however, does not account for the abrupt changes in slope at Cerro Azul, Sierra Negra, and Volcan Wolf, nor does it explain why Cerro Azul and Sierra Negra have such changes but not at the same elevation, though climatic conditions are similar.

It seems most probable that these great volcanoes had similar histories, characterized during earlier stages by greater accumulation of pyroclastic material which resulted in the development of steeper slopes. During later stages their history was chiefly one of lava accumulation. At Volcan Alcedo, the "fragmental" stage seems to have lasted longer and was more widespread than elsewhere.

Cerro Azul (Pl. 1, fig. 2), the smallest of the major volcanoes, has a deep crater with unusually steep walls and a well-marked, discontinuous bench about half a mile wide along its southwestern and western sides and two-thirds of the way up the rim from the deepest part of the crater. A smaller and obviously later crater appears in the southeastern part of the main crater. Marginal cliffs, especially along the northwestern, northern, and northeastern sides, show collapse of the crater floor, and one of these cuts directly across the smaller crater mentioned, which thus seems to have been fed at first from the marginal rift. There is a little evidence of fissure eruptions near this crater, but some recent lava streams flowed into the crater. Aerial photographs show many streams of dark, recent lava flowing northeastward from the northeastern rim and northwestward from the northwestern rim, the latter in part from lines of small cones that are elongated and breached almost wholly toward the north-northwest.

Sierra Negra (Plate 2), the lowest but widest of the five great craters, has a steep western wall marked by numerous rifts, but on the eastern and southern sides the crater rim declines more gradually and regularly to the floor. From its northern and northeastern rims the lava flows are directed radially, though those that flowed outward from its western edge were deflected northward off the steep slope rising southwest toward the higher Cerro Azul. This deflection suggests that Cerro Azul was both the earlier and the more rapidly growing

volcano, despite its smaller diameter. The great cone of Sierra Negra contains several recent flows, some of which appear from the rim as fresh surfaces, scarcely invaded by vegetation. Through valleys near the western edge (*see* Pl. 2, fig. 1) lava recently flowed. In that section of the crater high cliffs, now covered with lava fragments, run parallel to the crater rim, and indicate successive collapses of the crater floor with tilting; the more recent lavas erupted into and flowed down the resulting rifts. Such rifts are visible, though less distinctly, along the northern crater edge, where, on the outer slope, an area of small eruptive cones is called collectively "Volcan Chico". These cones are alined parallel to the crater rim and are sites of vapor emission (Pl. 2, fig. 2).

Volcan Alcedo is comparable to Sierra Negra in diameter of crater but is considerably higher. Inside its northern and northeastern walls a very well-marked bench, some 500 feet above the floor of the inner crater, conspicuously separates the higher and lower rimming cliffs. On the southwestern floor a prominent open cleft or rift trends northeastward. Hot springs and fumaroles are concentrated near this rift on the southwestern side of the crater.

Volcan Darwin (Pl. 3, fig. 1) is almost perfectly symmetrical. On its western slope the lowlands near the coast are mostly barren and consist of an apron, about 2 miles wide, and of flat-lying lava flows less than 100 feet above sea level, typical of those found elsewhere on the island. Several of these can be distinguished on the basis of color; they range from 15 to 25 feet in thickness. Their surfaces are ropy and scoriaceous and have numerous cavities up to 20 feet in diameter. Abundant V-shaped and essentially vertical tension fractures, up to 10 feet deep and 5 feet wide at the surface and hundreds of feet long, extend parallel to the slopes of the flows.

From sea level some of the flows on Volcan Darwin can be ascended to an altitude of 2500 feet. The slopes rise to 4000 feet above sea level in a distance of only 3 miles, an average angle of 18°, flattening slightly above 3000 feet. From the coastal apron up to 2500 feet, the vegetation includes trees as much as 20 feet high and a fairly thick growth in the gullies between flows. From 2500 feet to 3900 feet the cover is mainly brush, grass up to 3 feet high, and small trees. Some cinder is on the slopes here, but bare lava flows make up most of the exposed surface rock. Aerial photographs, such as Figure 1 of Plate 3, show the circular fractures marked by many craterlets around the main crater rim on the upper parts of the outer slope and the local extrusion of lava along such fractures. On the slopes near the crater rim the flows are generally less than 1000 feet wide, but many spread out to a mile in width farther down.

The crater rim of Volcan Darwin has an average width of some 1500 feet; its outer slope is pock-marked by many small cones 50 feet high and 50 feet in diameter, composed of scoriaceous lava. Much of the rim is covered with a stand of dense grass locally 8 feet high. By contrast, the floor of the crater is entirely barren of vegetation, which suggests the freshness of this lava. The crater itself, some 4 miles in average diameter, is almost circular. Its walls are very steep except on the eastern side, and 600–700 feet in height. Along the southeastern, southern, and southwestern base of these steep upper walls there is a flat bench from a few hundred feet to half a mile wide. The main crater floor, about 200 feet below this platform, appears to have subsided along a circular fault. On the eastern side of the crater, just inside the rim, a recent eruption from a small vent furnished lava that covers the eastern third of the main crater floor.

Volcan Wolf, the northernmost of the five volcanoes, was approached from the southwest (Pl. 3, fig. 2). From the slopes of this volcano great lava flows have poured all the way to the eastern shore of the island, and a few long ones came out on the northwest. Many flows streamed southward but were diverted southeastward and southwestward where they encountered the opposing slope of Volcan Darwin. An unusually well-marked rift runs about due north just within the southwestern edge of the crater rim and was apparently entered and flooded by flows from the crater slope to the west.

Lesser Craters and Cones

Darwin's estimate that the Galapagos Archipelago contains approximately 2000 lesser cones

and craters was proved modest by an examination of available aerial photographs covering about two-thirds of Isabela Island. In this part alone individual cones and vents number about 1600. From this an estimate of about 2500 cones may be inferred for this one island. Few of these exceed 500 feet in diameter and 200 feet in height. Some are cinder and scoria cones, consisting of coarse ejecta, others are spatter cones, formed by ejection and accumulation of molten lava, and still others are tuff cones. Many cones are complex. For instance, at Cerro Azul some of the numerous small cones within half a mile of the crater rim are made up entirely of scoria and cinders but others consist of a base of coarse cinder capped by a mantle, 1–3 feet thick, of fine-grained brown ash.

A typical volcanic field on the northeastern side of Sierra Negra contains many cones of the cinder and scoria type as well as spatter cones. The ejecta of the latter, appearing near each vent as continuous masses of lava, become thinner, finally being reduced to separate blebs of once-molten lava, resembling masses of brown, purplish-black, or black dough thrown out with force and fallen splattering upon the floor of older lava.

An especially striking pair of cones, each a little over a mile across, may be seen on the western shore of the island, southwest of Volcan Darwin (Pl. 1, fig. 1). These are locally called Beagle and Darwin, but in view of the great crater previously described and officially named Darwin, they might better be called Tagus (to the north) and Beagle (south) after Chubb (1933, p. 12). Both cones, made up chiefly of fragmental material, are breached on their southern sides, and the northerly crater has been invaded by the sea to form Tagus Bay.

Remnants of other, larger craters are also present on the island. One comparable in dimensions to the five main craters already described, is at Cabo (Cape) Berkeley, near the end of the peninsula extending west from Volcan Wolf. Only the eastern wall stands today; the western side was breached by explosion, or dropped by faulting, or truncated by the sea. From this center flows extend northeast and southeast to the coast. Remnants of another great crater form Iguana Bay on the southwestern shore; one of the related marginal rifts is exposed along the shore. This crater-rim fragment, like that at Cabo Berkeley, is marked by steep cliffs, several hundred feet high, fronting the sea and by a shelf typical of the inner shelves of the five main craters. Thus Iguana Bay has both the plan and the topography of a crater breached to the southwest. Most of the lesser islands surrounding Isabela, especially along the southeastern shore (Fig. 2), are breached to the south; examples are Tortuga Island and the three Hermanos (Crossman) Islands. Several other breached craters are suggested by incomplete soundings, but even the highest points of these no longer rise above sea level.

Fissures and Rifts

On Isabela many of the longer flows that contributed so much to the height of the major cones apparently originated from separate vents around the present rims of the craters or possibly from vents now destroyed, at even higher elevations on the great cones. Indeed, on Isabela there is no clear evidence that any great quantity of lava came out over the present rims of the craters at the levels at which they now stand, unless it be inferred that there has been a great subsequent depression of the crater floors.

However, many of the flows here as elsewhere were from fissures. The wide distribution and great importance of such fissures are conspicuous on Isabela. Many of the dark-lava streams begin uphill along sharp lines on coneless slopes —the lines being traces of fissures. In some cases other flows originated some distance away along the same line. The area west of Volcan Wolf and east of the relict volcano at Cabo Berkeley shows perhaps most clearly eruption from fissures. An outstanding illustration of this is on the southeastern slope of Volcan Wolf. On aerial photographs some flows give a false impression of being fissure flows by entering trenches or rifts, filling them, and then flowing out again farther downslope as does one group of flows in the southwestern part of the outer crater wall and the floor of Volcan Wolf (Pl. 3, fig. 2).

Many fairly well-defined fissures on the lower slopes of the great cones were not filled by the longer fingerlike flows that came from above, for the aerial photographs (Pl. 4, fig. 2) show

the fissures still intact and continuous, though within the boundaries of these flows.

One of the most graphic features of the five great volcanoes is the suggested presence of a system of fissures radiating from each master crater; these apparently controlled the distribution of parasitic cones on the slopes. North of the northern rim of Volcan Wolf the effect of these fissures can be seen very clearly (Pl. 4); lesser craters and especially cinder cones lie in almost continuous lines, like beads radially arranged with regard to the main crater. Here the more easterly cinder cones are elongated toward the northeast, aligned in the same direction, and breached, for the most part, on their northeastern sides. Farther west (Pl. 4, fig. 1) they are elongated northward and breached, if at all, in their downslope, northern margins. Still farther westward along the northwestern part of the crater rim, they trend outward successively west-northwest (Pl. 4, fig. 2), then west-northwest by west. Moreover, such craterlets are in reality not circular in plan but ovoid, canoe-shaped, or even boat-shaped, or they assume the form of twin craters, the downslope edge alone being breached; or this edge is the more deeply breached, whereas the upslope side is lightly breached and the gunwales, as it were, are still intact. This radial elongation of some of the cones and the parallel alinement of fissures and small cones is most marked midway down the slopes of the great volcanoes. Farther down, where the slope is less, the little cones tend to become more circular, the controlling rifts are less visible or absent, and where they are present their long dimensions are less accordant with the radial pattern. Alinement is especially conspicuous in areas where the surface rock is pyroclastic and recent flows are not extensive. Where flows are the chief cover, the craters and their surrounding accumulation are more nearly circular. This is to be expected because pyroclastics are more easily breached, and such areas therefore reflect the fissure pattern more exactly.

The presence below the surface of another type of rift having circumferential rather than radial directions is also suggested on Isabela. Similar cases have been cited by Cotton (1952, p. 219–220). Such rifts are marked on Isabela not by elongated depressions or visible rifts but by the alinement of small, essentially circular cones. The circular arrangement of these cones is especially noteworthy around the five great craters (Pl. 1, fig. 2; Pls. 2, 3), and some of them are accompanied by solfataras, as strikingly illustrated on the northern side of Sierra Negra (Pl. 2, fig. 2). The general trend of the lines of these craters in the locality mentioned is due east, that is, circumferential—parallel to subsidence fractures on the north crater wall; even here, however, elongated craters lie in the radial direction and in directions oblique or normal to this. Similar circumferential crater patterns appear on Cerro Azul, especially on the northern and southwestern margins, half a mile outside the crater wall; on Volcan Wolf, crossing the eastern and southeastern slopes near the crest, and again half a mile outside the main crater; in several places on Volcan Darwin; and forming a well-defined line some 8 miles long, which, though not related to any existing crater, is clearly traceable from Cabo Berkeley eastward toward Volcan Wolf, on the southerly slope of which the line appears to be buried in lavas from the great volcano.

Almost all fissures governing the explosions are therefore either radial or circumferential. They probably represent areas of stretching ("tumescence") of the chilled surface of the cone with upward thrust of the covered lava; this explanation would account for their resemblance in pattern to tension fractures.

In addition to the narrow fissures or rifts mentioned, there are here, as in the Hawaiian volcanoes (Wolff, 1914, p. 457–458) and elsewhere, other visible fractures or depressed areas within the craters apparently caused by subsidence of the crater floors. A rift of this type, trending in a zigzag course generally northeast and ranging in width up to 100 feet, appears on the floor of Volcan Alcedo along its southerly and southeasterly side; to the southeast, between it and the crater rim, are many solfataras and one major hot spring. A similar rift is on the southwestern edge of the crater floor of Volcan Wolf; into this at some recent time lava flowed from the cliff above (Pl. 3, fig. 2). A complex set of such rifts probably runs parallel to the west wall of Sierra Negra, inside the crater, as shown by steep slopes on both sides (Pl. 2, fig. 1). This subsidence trough, however, is

today filled by a younger stream of lava. Such subsidence may likewise account for the steep, sinuous ridge east of the sulphur deposit in this region.

can Alcedo, and the northeastern wall of Volcan Wolf) indicate subsidence of this kind by higher crater-floor levels.

The pattern of flows and rifts is indicated on

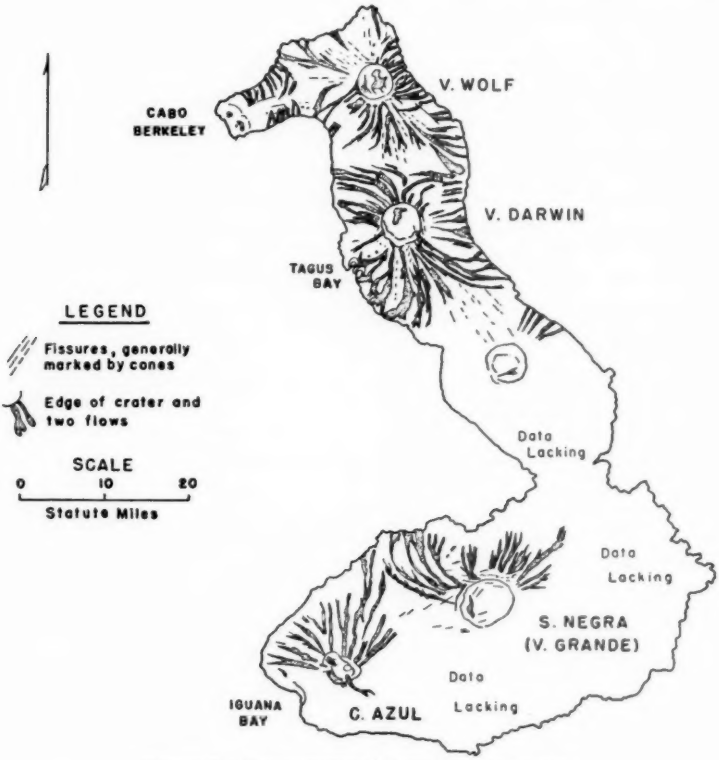

FIGURE 3.—DIAGRAMMATIC MAP OF ISABELA ISLAND
Showing known rifts and flows; from field study and aerial photographs.

Many such rifts are flanked by evidences of even more-widespread subsidence of the crater floor. In the crater of Sierra Negra, west of the lava stream filling the rift just described, steep cliffs are visible from the floor of the caldera but not from the air; they suggest facet spurs or earthquake scarps. Several little cones are located in or near these cliffs and spurs. Above the scarps are stretches typical of landslide topography—the steeper slopes face craterward, and the surface, originally essentially flat, is now tipped backward, that is, toward the west, away from the crater. The benches within the inner edges of several craters (Cerro Azul, Volcan Alcedo, and the northeastern wall of Volcan Wolf) indicate subsidence of this kind by higher crater-floor levels.

The pattern of flows and rifts is indicated on Figure 3, assembled from a study of aerial photographs and field observations.

CAUSES OF CRATER BREACHING

Darwin was impressed by the breaching of many Galapagos craters toward the south, attributing it to marine erosion, intensified by open-ocean swells and the trade-wind direction (1839, p. 358). To the present writers a different explanation seems more probable, at least for the breached craters on Isabela Island.

Few of the continuous fissures that accompany the alinement and breaching of the sub-

sidiary cones, whether radial or circumferential, could be seen to extend to a depth of more than 10–15 feet.They are wider above than below, and their walls are essentially vertical. They are best interpreted as tensional, rather than compressional features, the more so as many of them are filled by later up-welling lava, the basic "dikes" described later. These recent fractures must be developed when the five great cones already stood with their upper parts from 3000 to 7000 feet above their sea-level or submarine bases. Hence, lateral compression could hardly have been transmitted through their surface rocks. Such considerations make it most unlikely that the surface fracture pattern here described can be attributed to lateral stresses, acting tangentially through the crust of the earth, as has been suggested for similar patterns studied elsewhere by Anderson (1951, p. 31–57) and Ode (1953).

The upward thrust of the lava rising repeatedly within each crater probably stretched and tensionally fractured the crust over the conduit. Thus the lava could move not only upward along the concentric fractures around the volcanic throat, but also laterally, following the radial fissures. As the doming increased or was renewed, new cracks developed from the older ones, and the older radial ones were lengthened. Hence there would be a tendency for the lava to thrust not only upward from the depths against the cover of already chilled lava; it would also press outward from the main conduit. Thus a given craterlet or spatter cone would be breached at the side farthest from the mother crater. This obliquely upward thrust would be favored also by the lighter load of rock on the downslope side of the crater.

This explanation is believed to account for the direction of breaching of the myriad small cones on the slopes of the great shield volcanoes. In general, too, it may explain the southeastward breaching of the partially submerged crater islets marginal to Isabela, if their position (usually to the southeast) in relation to the nearest large shield volcano is considered.

Description of Flows

Some lava flows of Isabela Island have smooth surfaces, like those making up the great part of Volcan Alcedo, where, however, the flows along the eastern rim, though petrographically similar, are made up largely of a flow breccia of angular blocks. Other flows are so rough that they are dangerous to walk across because of the sharp thin edges of their fragments. On the northeastern edge of Sierra Negra, for example, lathshaped lava fragments no more than 4 inches wide and half an inch thick rise at all angles as much as 3 feet above the rest of the flow.

Basic "dikes," consisting of lava like that through which they cut, transect recent lava flows and can be followed several hundred feet in places where the lava surface is well exposed. These are virtually fissure eruptions chilled at the surface before they had flowed from the source aperture. Examples were seen cutting the smoother flows on the floor of Volcan Alcedo and in the Volcan Chico area just outside the crater on the north rim of Sierra Negra. Where they occupy the curving and concentric marginal fissures of craters just mentioned, such "dikes" resemble "ring dikes" in pattern, as do some of the curving fissures along the western wall of the Alcedo crater. Similar "dikes" have been described from other localities (Gevers, 1941, pl. 31).

Slickensided walls suggest the highly viscous state of this molten rock, subject to enough internal friction while partly solid to allow internal striae to develop. On the north side of Sierra Negra a block 7 by 8 feet was found cut by a series of parallel scorings, the deepest groove with an amplitude of 2 inches; the grooved surface dipped in the general direction in which the flow moved, as determined by its general slope, and the striae ran directly down the dip.

Recent Eruptive History

None of the five main volcanic centers has experienced a major eruption since discovery, but minor eruptions were reported in 1814, 1825, and 1897, and aerial photographs indicate several relatively recent ones. In 1925, Beebe (1926) observed an eruption of lava in the vicinity of Volcan Wolf and Volcan Darwin. Since 1934, several outbreaks have occurred on Cerro Azul, in the southwestern part of the

island—the latest in 1951; the lava from one of these blocked the trail leading from Cerro Azul to Villamil, and lava from another (1949?) covered or destroyed, probably by volatilization, some sulphur deposits once worked in the crater of that volcano. An eruption of Sierra Negra in 1948 discharged enough fine material to be noticeable at Villamil but did no damage even to near-by Santo Tomas, partly because the wind blew away from this settlement at the time. Recently Volcan Wolf erupted violently, the first reported activity being on January 24, 1948, according to Joseph Madruga, as described by Richards (1954). This came from a small cone, some 100 feet high, on the southeastern side of the main crater, at an altitude about 4000 feet above sea level.

The eruption on Isabela last recorded was in August, 1953 (Richards, 1954). On August 27 and 31 activity was described by radio messages from three different ships. It consisted mainly of a fissure eruption ejecting both fragments and liquid and located on the north-northwestern side of Sierra Negra, apparently in the neighborhood of Volcan Chico. The source fissure was at an elevation of 3700 feet, probably trended northeast, and was about 2 miles long. At least one flow extended northwest from it to the sea, a distance of some 10 miles, and elsewhere three shorter flows came from the same fissure. Masses of volcanic ash were periodically blown into the air. Activity of this general nature continued to January, 1954, the last date of recorded observations.

Volcanic Sequence and Age

On the five great craters, the chief rock below altitudes of 2000 feet is lava. In view of this preponderance and of the pyroclastic deposits higher on the rims, there was apparently an earlier stage mainly characterized by lava accumulation and building of the great craters, followed by a later stage in which pyroclastic material predominated. However, the greater part of the very fresh, lesser flows now visible at the surface are later than much if not most of the pyroclastic material. Thus, the western slope of Volcan Wolf, above 2000 feet, seems to be made up almost wholly of clastic material which partly buried earlier flows of lava; nevertheless, even here the evidence as to time relations is not conclusive.

Many of the smaller, gray cones of pyroclastics on the slopes of the five great volcanoes lie like islands immersed in more recent lava flows. Most of these flows stem from the last great eruptive epoch. Many flows—for instance, those on the northern slope and inside the western crater wall of Sierra Negra—are very recent, as shown by their fresh surfaces and the lack of plant cover. Such lavas flowed across the scoria-covered landscape that makes up so much of the slopes of Mounts Darwin and Wolf. Again, on the eastern shore of Isabela the lowest rock exposed is a black basalt at tide level; this is followed immediately by several hundred feet of scoria at the base of which is a distinctive highly porous pumiceous flow. Such clastic ejecta were later partly covered by several distinctly fresh flows of more basic lava that poured across them to the sea. Similarly, on the west coast, the ash and scoria cones of Beagle and Tagus craters were surrounded and Beagle crater was actually invaded by late lava.

In brief, these observations support qualitatively the conclusions of Wolf (1895, p. 249) and Chubb (1933, p. 7, 12, 16, 22) that the major fresh flows visible are generally later, the clastic ejecta earlier in the histories of the great volcanoes on Isabela. It is of course possible and indeed probable that the prominent clastic deposits near the tops of the cones were interbedded with considerable masses of still earlier flows. The upper slopes of the great cones and the major exposed rocks of the partially submerged islands off the southeastern shore of Isabela Island may represent the dominantly clastic stage in the local volcanic history. It seems clear, however, that neither stage was confined solely to one type of material.

Little is known concerning the age of the volcanic rocks of Isabela Island. Ochsner (Dall, 1924) collected fossils on Santa Cruz (Indefatigable) Island and the adjacent Seymour Islet just north of Santa Cruz, and these were tentatively regarded as lower Pliocene. Another shell bed was found by Ochsner between lavas 40–50 feet above the ocean near Villamil on Isabela Island. These appear to be "old Pleistocene," but Dall (1924) noted that 50 species in the fauna are still living in Panama. The com-

plexion of this fauna is American, not Indian. The evidence suggests that the islands and their volcanism are young and that the more easterly part of the archipelago is relatively the older.

Petrography

With rare exceptions the rocks well enough crystallized to be clearly classifiable are all very similar. Most are porphyritic basalts with microcrystalline groundmass and phenocrysts, chiefly of basic plagioclase. The phenocrysts form between 2 and 20 per cent of the volume. Many lavas are scoriaceous and many show some alteration. In general, the excellent descriptions given by Richardson (1933, p. 51–54) fit the cases studied very well. In composition and texture the scoriaceous basalt mentioned by Richardson from Tagus Cove is strikingly similar to lava collected during the present studies on the eastern shore just east of Volcan Alcedo. The rock on the eastern shore is a rather fine-grained basaltic or basaltic andesite, still quite fresh. Plagioclase phenocrysts (An_{70}) up to 1 mm in diameter make up some 8–10 per cent of the total volume, vesicles perhaps 10–12 per cent. The groundmass consists of tiny feldspar laths fluidally alined, having a composition of about An_{55} and forming some 40 per cent of the total volume; magnetite in numerous tiny grains and elongated shreds forms 10 per cent; many small very light-yellow augite crystals make up 10 per cent; olivine is extremely rare; and a very light-brown glass is evenly distributed. This rock suggests also the basalt porphyry with plagioclase of Lacroix (1927, p. 68–69).

Similar rock forms the major flow with which the sulphur is associated in the crater of Sierra Negra. The mineral content is almost identical and so is the texture, except that the phenocrysts of labradorite are fewer and the glass is more commonly aggregated into areas comparable to the phenocrysts in the rock described from the eastern shore.

Analyses of rocks of this type are characterized by low K_2O and high Al_2O_3 and CaO content, like those described by Richardson (1933, p. 64). Analyses of these specimens and similar published analyses are given in Table 2.

One exceptional specimen is a scoriaceous lava from the east coast, about due southeast of Volcan Alcedo, which is almost a pumice, very pale in color and very frothy. Another is a grayish rock, lighter in color than those described and bearing more phenocrysts, so much so that this rock approaches a granite texturally. It occurs in loose blocks apparently fallen from the cliffs rimming the northeastern wall of the crater of Volcan Wolf. Under the microscope about 30 per cent consists of coarsely crystalline anorthite (An_{80}), a small amount (4–6 per cent) of light-green augite, and a yet smaller quantity (2–3 per cent) of shattered olivine. These together make up the phenocrysts, which are embedded in a groundmass 25 per cent of which consists of laths and needles of labradorite (An_{60}), 20 per cent of lenses and interstitial fillings of yellow to light-brown glass, and 15 per cent of magnetite in angular shreds and irregularly shaped grains. Despite its coarse texture, lighter color, and more conspicuous content of pyroxene and olivine, this rock conforms to the type generally found on the island. It is perhaps significant that this distinct variety occurs, however, in the northern part of Isabela Island, not very far from the high-olivine basalt area of James Island, described by Richardson (1933, p. 48–49).

In brief, all the rocks studied are basalts, almost andesitic, with plagioclase as the dominant constituent, high in magnetite and relatively low in pyroxenes and olivine. There are basaltic tuffs and agglomerates as well as flows. Of the occasional trachyte, unequivocal andesite, and spilite mentioned by Richardson from elsewhere in the Galapagos Islands (1933, p. 46–48) no trace was found on Isabela.

Megatectonic Features

Major Lines of Volcanicity

Darwin (1890, p. 249–250) suggested that in the archipelago the main crater lines in the Galapagos Islands were two-fold: east-northeast and north-northwest; this conclusion was criticized by Wolf (1895, p. 252), but has since been supported by Chubb (1933, p. 21) and others. The southern part of Isabela is alined east-northeast, the northern part north-northwest. From a study of the U. S. Navy's Hydro-

graphic Chart No. 1789 (1946 ed.), however, the other islands appear far less clearly elongated except for San Cristobal and, to a lesser degree, San Salvador, Española, and Pinta. Most of the islands show a massed cluster of craters, rather than a pronounced alinement of several.

More careful study of the islands themselves and of soundings in the intervening channels, however, supports the concept of Darwin's two directions (Fig. 4). The Canal de Isabela (Fig. 2), separating the east coast of Isabela Island from the Island of San Salvador at Punta Boquerizo, Isola Rabida, and the Island of Santa Cruz, is matched by an upland to the east, consisting of Santa Cruz, San Salvador and the shallow sea between. South-southeastward the Canal de Isabela ends, first by shoaling, then by the actual rise of a volcanic island, Pinzon, approximately in the line suggested by Cerro Azul and Sierra Negra craters on Isabela; the eastward continuation of this line would pass 15 miles north of the twin peaks of Santa Cruz Island. A line almost parallel to that just sketched can be drawn from the magnificent peak of Fernandine (Narborough) Island through Volcan Darwin (Fig. 2) and a similar line through Volcan Alcedo and the highest points of San Salvador. The nearly parallel line of the northern lobe of Isabela, from Cabo Berkeley through Volcan Wolf has already been commented upon. These are perhaps the best-defined volcanic alinements.

Farther east the east-northeast trend, very gently concave to the northwest, may be represented by the alinement of Santa Maria (Floreana) Island, the shoals of Banco Hancock or of Arrecife Macgowen, and the markedly elongated San Cristobal (Chatham) Island.

TABLE 2.—CHEMICAL ANALYSES OF ROCKS, ISABELA ISLAND

	No. 1	No. 2	No. 3	No. 4	No. 5
SiO_2	47.01	47.12	47.80	48.30	48.10
TiO_2	3.20	3.44	2.40	2.99	1.90
Al_2O_3	15.57	14.93	18.31	13.35	15.90
Fe_2O_3	2.32	2.57	1.47	4.76	1.93
FeO	11.57	13.06	8.20	10.33	10.28
MnO	0.20	0.27	0.14	0.19	0.29
MgO	5.25	5.25	4.89	5.30	6.28
CaO	9.77	9.44	13.00	9.60	11.60
Na_2O	3.00	2.57	2.48	2.46	2.68
K_2O	0.31	1.01	0.57	0.77	0.30
P_2O_5	0.32	0.36	0.41	0.25	0.23
H_2O^+	1.40	0.36	0.31	1.75	0.40
H_2O^- (110°C.)	0.24	0.01	0.11	0.19	0.10
CO_2	0.00	0.00	n.d.	0.00	n.d.
S	0.06	0.04	n.d.	0.09	tr.
SO_3	0.46	0.15	n.d.	0.24	n.d.
Cl	0.08	0.09	n.d.	0.05	n.d.
Total	100.76	100.67	100.09	100.62	99.99

1. Scoriaceous plagioclase basalt, unaltered, from Sierra Negra near westernmost solfatara area, west side of crater; collected by Banfield, analyzed by H. B. Wiik, Helsingfors.

2. Plagioclase basaltic tuff, unaltered, from Sierra Negra, near easternmost solfatara area, south central part of crater; collected by Banfield, analyzed by H. B. Wiik, Helsingfors.

3. Plagioclase basalt, scoriaceous basal lava, Tagus Cove, collected by Lacroix, analyzed by Raoult, republished by Richardson (1933, p. 52).

4. Plagioclase basalt, unaltered, from Sierra Negra near westernmost solfatara area at western side of crater; collected by Behre, analyzed by H. B. Wiik, Helsingfors.

5. Olivine andesite, ejected fragment; analyzed by Herdsman, published by Richardson (1933, p. 52).

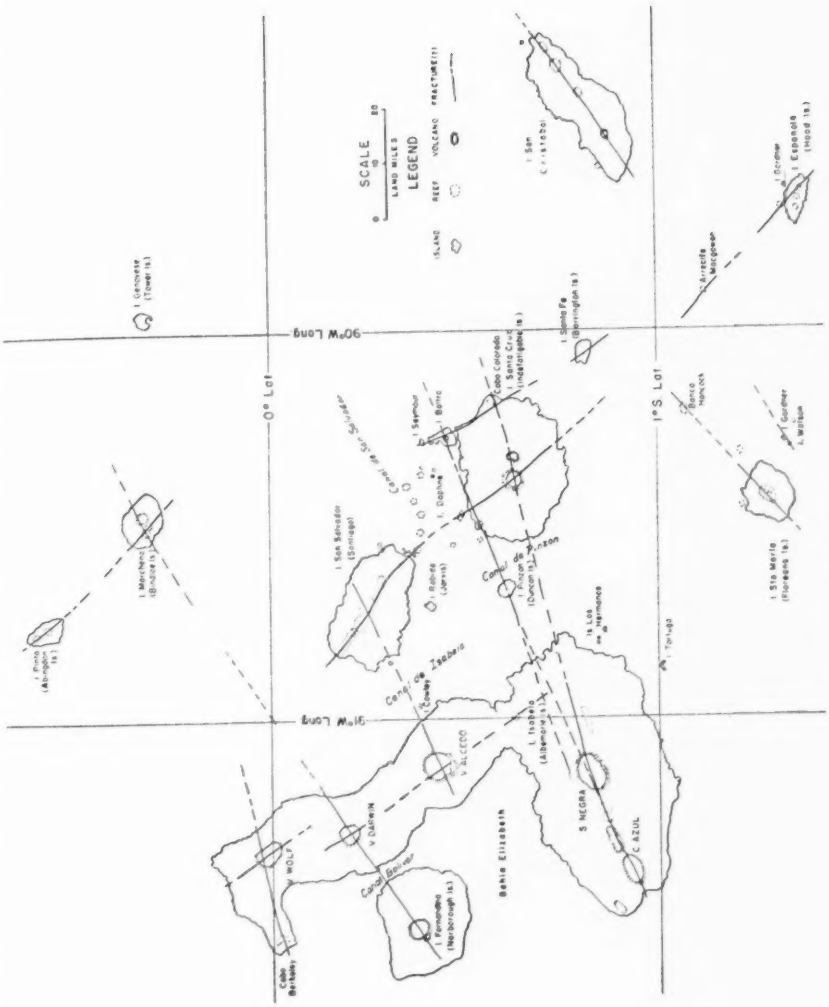

FIGURE 4.—MAP OF GALAPAGOS ARCHIPELAGO
Lines of hypothetical major fractures are shown as inferred from positions of volcanic cones

Other north-northwest lines are suggested by the islands Pinta (Abingdon) and Marchena (Bindloe) and the shoals in this line southeast of Marchena Island. The southeastward continuation of a line nearly parallel and almost superimposed upon the San Salvador-Santa Cruz line may be represented by Banco Hancock and (or) Arrecife Macgowen and finally, despite its apparently opposite elongation, Española (Hood) Island. A line nearly parallel with that mentioned extends through Baltra Island, Cabo Colorado on Santa Cruz, and Santa Fe (Barrington) Island.

In short, Darwin's hypothesis of alinement seems feasible and the further suggestion may be entertained that the deeper vents are controlled by fissure intersections. The radial fissures governing the location of the minor vents on the slopes of the great volcanoes are not believed to be related to megatectonic stresses, however.

Continental Connections

An even larger problem bearing on the peculiar biota of the archipelago is, were there once connections with the mainland of Central America, and if so, where did the connections lie? Finding fossiliferous beds of early Tertiary or greater age would be a great help but this is most unlikely. Soundings, if accurate and closely spaced, would be invaluable, but the bottom topography between the mainland and the islands is only roughly known. What bathymetric lines are available from the maps of the Bureau Hydrographique International of Monaco (1949) favor the presence of a "plateau" at a depth of about 3000 m, extending from Costa Rica southwest through Cocos Island to the Galapagos (Fig. 1), as suggested by Beebe (1924, p. 418–429), who correctly describes the relations between the islands and this "plateau." As the Cocos and Galapagos rocks are all volcanic and chiefly basaltic and as the configuration and therefore the origin of the "plateau" are still not clearly understood, there is no geologic proof that these islands were once part of a landmass connected with the continent either at Costa Rica or farther south. Nor is there strong evidence favoring the inference of Baur (1897) of a once-continuous landmass, later dismembered into islands by subsidence.

There is, however, some suggestion that the archipelago is advancing westward (Chubb, 1933, p. 21–22), for volcanicity is most active today on Isabela and Fernandina. Moreover, prehistoric volcanic features are fresher in the western islands of the group, though they may merely reflect differences in climate and hence in rock weathering. Most probably the building up of volcanic piles from the ocean floor has moved successively southwestward while erosion planed off the eastern islands as vulcanicity declined there. If so, the continental aspects of the terrestrial biota on the islands and the entire problem of "seeding" the islands with land life might be explained by the growth and later marine erosion of a succession of stepping stones, developing westward, their eastern bridgelike members having later been largely destroyed.

Vent alinement and progressive migration of active vulcanism in a constant direction were described in a specific example by Easton (1930, p. 99) and summarized by Cotton (1952, p. 216–220), who stressed the relation of migration to such alinement. Whether this principle operated in the Galapagos, however, can at best only be deducted. For a long period discussions of the location of Vesuvius by Johnston-Lavis (1884) and Perret (1924, p. 13) turned in part on whether localization was fortuitous and along a single fracture or instead at the intersection of two fractures; Perret, in contrast with Johnston-Lavis and most earlier writers, supported the later, single fissure theory. Both hypotheses favor progressive migration, if migration there be, toward the deeper sea, whether in downwarps transverse to the coast or downdip on monoclinal seaward-dipping structures. This direction of migration is in accord with the proposed relative ages of the volcanoes of Galapagos.

In his clear summary, Chubb (1933, p. 43–44) suggests that the Galapagos and the Marquesas islands may be at the opposite ends of a plateau —the Albatross plateau—of which Cocos Island, northeast of the Galapagos and characterized by alkaline rocks, may be an outlier. For this viewpoint the combined data of soundings and petrology are fairly strong evidence, though not conclusive. The volcanoes here described would rise progressively southwestward, out of this plateau.

Solfataras

Location and Types

Solfataras, discharging steam or steam and sulphurous emanations, occur in the craters of three of the five main volcanoes—Sierra Negra, Volcan Alcedo, and Volcan Wolf—and on the northeast flank of Sierra Negra. Such solfataras have been formed and others have been destroyed within the last few years. The solfataras may be classified for convenience into two general types—sulphur-bearing and sulphur-free. The former are described in some detail because of their possible economic interest. Sulphur-bearing solfataras were reported on Cerro Azul about 1934, but these apparently were destroyed by recent volcanic eruptions; neither sulphur deposits nor related phenomena were seen there in February 1953, except for one small steam vent on the south rim.

Several solfataras are on the northeastern wall of the crater of Volcan Wolf at an altitude of 3800 feet above the sea, 1300 feet below the crater rim, and 200 feet above the crater floor. Here two areas containing sulphur solfataras are about 400 feet apart and each is approximately an acre in extent.

Along the southwest wall of the crater of Volcan Alcedo at altitudes between 2640 and 3250 feet is an area containing both steam and sulphur solfataras with local concentrations. This sulphur area extends about 2½ miles with a maximum width of half a mile.

At Sierra Negra three areas contain sulphur-bearing solfataras. Two lie within the crater near its western side at an elevation some 3000 feet above sea level (Pl. 2, fig. 1). The third is situated on the northeast flank of the crater, some 4000 feet from the rim, at an altitude of about 2850 feet (Pl. 2, fig. 2). Locally this latter area is called Volcancito or Volcan Chico, though the name is not confined to any specific crater. In the westernmost of the three Sierra Negra localities, solfataras occur over a surface half a mile long and 700 feet wide; in the southernmost they are scattered over an area 1000 by 500 feet. At Volcancito the solfataras appear in a band some 3000 feet long and 500 feet wide. No other solfataras bearing sulphur or steam in quantity were observed on Sierra Negra, but this large crater was not completely explored.

Description

Perhaps the coolest and largest of the nonsulphurous solfataras was found at the southern end of the solfataric area in the crater of Volcan Alcedo. Here, half way up the side of the crater, is a basin about 100 feet in diameter filled with water to within 50 feet of its brim and almost boiling. Large bubbles of steam continually rise to the surface of the water. There is no evidence of encrustations or alteration around the basin; heavy vegetation grows to the water's edge.

Field evidence indicates that the main gas from this nonsulphurous solfatara consists wholly of water vapor issuing from small holes or cracks in lava. With rare exceptions, such gases appear not to have reacted strongly with the country rock which is at most only very slightly bleached or limonite-stained. In some localities, as on Volcan Alcedo, hot water preponderates over steam and fills shallow basins. There is no deposition from this type of solfatara. The volume of steam emitted increases very noticeably after a heavy rainstorm evidently because of an increase of inflowing rainwater. The nonsulphurous solfataras are found in and around areas of sulphurous solfataras and are also widely distributed as isolated vents in the floors of the craters.

The sulphur-bearing solfataras are more closely grouped than the nonsulphurous type, which suggests a higher temperature. All three sulphur-bearing areas on Sierra Negra and those on Volcan Alcedo and Volcan Wolf contain deposits of sulphur originating from gases emitted from abundant small vents. At Sierra Negra several larger vents occur at different levels of a series of flows.

Sulphur-bearing and nonsulphurous solfataric areas alike generally appear along local lines of weakness, roughly parallel to the rims of the craters. All observed cases of sulphur deposition were just within or outside the margins of the craters, along or very near areas of subsidence, rifts, or related fractures. In such areas an acre may contain a hundred or more small sulphur-encrusted vents. Here sulphur deposition is continuous.

The sulphur vents themselves are usually along recognizable joints or other fractures and are strikingly alined. They are noticeably rare in ropy or scoriaceous lava and far more common

and well developed in the massive, more jointed flows and in cinders. A typical sulphur vent rises half a foot or less above the general level and may have an opening 6 inches or less in diameter, around the rim of which is an area of sulphur concentration that, in the case of a small vent, extends radially a foot or so. In some instances the rims of such vents are built up in the form of pipes as much as a foot in height by deposition of almost pure (96–98 per cent) sulphur. The walls of these pipes are generally less than an inch thick.

Around the sulphurous solfataras the ground is hot. At a depth of 3 or 4 feet, the ground temperature is generally about 125° F, and the fresh rock or altered surface material is too hot either to stand on or to handle. Much of this heat probably comes from oxidation of the rising gases, as the country rock only 50 or 100 feet away commonly has a normal temperature at the surface.

The gases emanating from the sulphurous vents are generally very pale yellowish in color and consist partly of sulphur but apparently mainly of steam. The odor of hydrogen sulphide is readily recognizable when the air is dry. No apparatus was available to make field tests for other possible ions. Immediately after a heavy rainfall, the odor of the emitted gas changes noticeably to the acrid one of sulphur dioxide.

In places, the areas of sulphur deposition are increasing in size. For example, in Volcan Alcedo there are many dead trees and other plant skeletons in the sulphur-bearing area, indicating that the vents and sulphur deposition are of recent origin. It is possible, of course, that the volume of gases being emitted has not changed, but that the solfataric vents have shifted to new localities. Thus, in one of the areas within the crater of Sierra Negra a very active solfatara is about 100 feet away from the main area. Native guides stated that this vent is less than 3 years old.

The deposition of sulphur and formation of the whitish, clay-like material derived from the alteration of the original rock form a somewhat impervious capping at the surface that in time diverts the gases to other channels of easier escape. As most deposits are situated on slopes, this effect is reduced by the erosive action of heavy rainfall, which removes the loose or unconsolidated material.

The sulphur deposited from the gases is very pure. A characteristic sample from one of the little "pipes" surrounding a vent contained 98.61 per cent sulphur, 0.001 per cent arsenic, and 0.002 per cent antimony. The remaining constituents were not determined, but probably consist mainly of moisture and dust.

Where the topography is favorable and there is a prevailing wind, the sulphur sublimes in decreasing amounts for considerable distances from the solfataras. For example, at the main deposits in Sierra Negra the rocks to the north of the deposit are encrusted with sulphur for a distance of 1000 feet, the maximum thickness being about a quarter of an inch. This sulphur is greenish and apparently has associated with

PLATE 1.—AERIAL PHOTOGRAPHS

FIGURE 1.—BEAGLE (LOWER, SOUTH) AND TAGUS (UPPER, NORTH) VOLCANOES, WESTERN SHORE, ISABELA ISLAND

Lavas are dark, cinder and ash are light; photographed from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet.

FIGURE 2.—CERRO AZUL

Photographed from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. *Note* two sets of concentric subsidence cliffs, with bench between, and subordinate cone in northeastern part of main crater.

PLATE 2.—AERIAL PHOTOGRAPHS OF SIERRA NEGRA CRATER

FIGURE 1.—WESTERN AND SOUTHWESTERN PART

Photographed from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. White spots on west and south are solfataras. *Note* zig-zag subsidence cliff; also dimples of craterlets parallel to margin outside main crater.

FIGURE 2.—NORTHEASTERLY PART

Photographed from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. View partly obscured by clouds. *Note* small cones, alined along otherwise poorly marked rifts; also bleaching and distributive subsidence along north rim of main crater.

FIGURE 1

FIGURE 2

AERIAL PHOTOGRAPHS

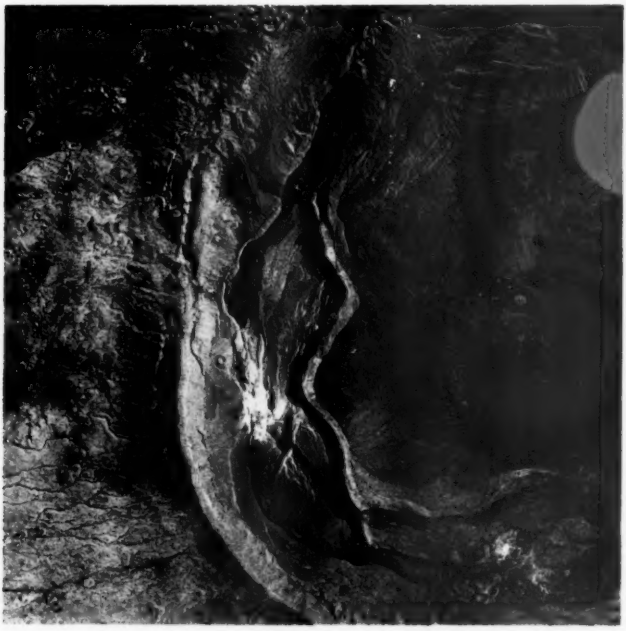

FIGURE 1

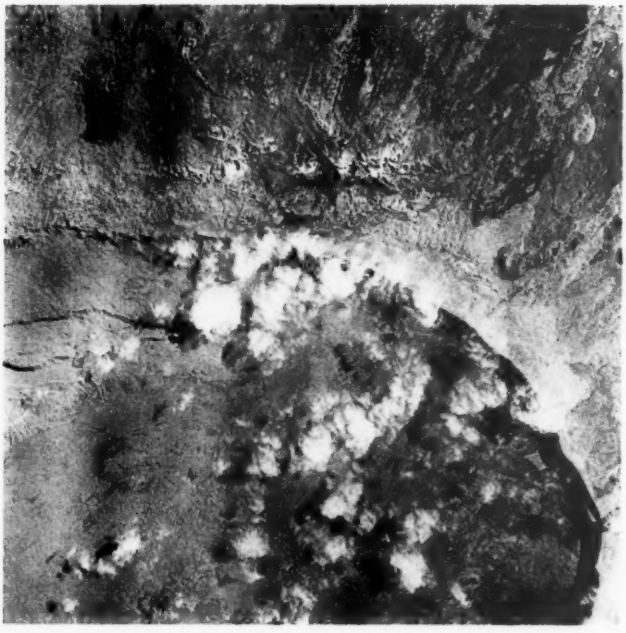

FIGURE 2

AERIAL PHOTOGRAPHS OF SIERRA NEGRA CRATER

FIGURE 1

FIGURE 2

AERIAL PHOTOGRAPHS

FIGURE 1

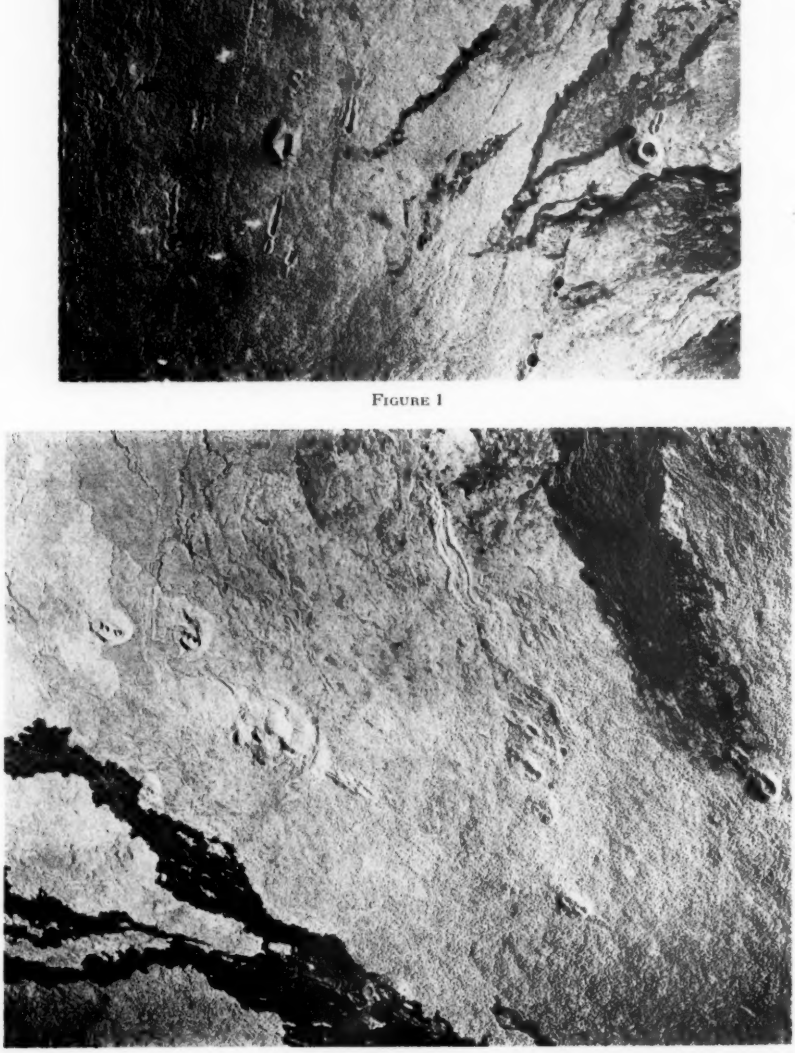

FIGURE 2

AERIAL PHOTOGRAPHS OF VOLCAN WOLF

it some sulphur-bearing soluble substance, presumably a sulphate. Sulphur of similar color cements rock fragments in the alluvium below the deposit.

Rock Alteration near Sulphur Vents

Characteristic of the areas around such sulphurous vents is intense alteration of country rock from hard, dark lava to a white claylike material which, in the final stages of alteration, is a mud, very sticky and plastic when wet. A recent study, similar to this one but where lavas were acidic rather than basic, is reported by Steiner (1953). Such areas, because of their light color, may be identified on aerial photographs (Pl. 2). Where three dimensions are exposed, as along steep slopes, the smaller areas of alteration tend to be shaped like cones, with a vertical axis and a circular outline at the surface, a pattern similar to that of the sulphur-rich areas themselves. Such areas, like those of sulphur deposition, are generally alined as though along fissures.

Intense alteration, like that on the rock benches immediately adjoining the western solfatara in Sierra Negra, has proceeded along joints. The successive steps are recognizable. The fresh dark-gray rock, jointed like most lavas, shows thin light-gray to white bands on both sides of the joint. These widen as the center of alteration is approached. Next the bleached rock marginal to the joints becomes soft and clayey and expands so that the joints are obscured or disappear. At this stage vesicles, if present, may still be visible, and though the rock becomes pale gray or creamy as a whole, its igneous texture persists.

While this general bleaching and alteration spread into the rock, the unjointed areas become darker as though the iron removed by the bleaching migrated in from the rock margins. Such altered rock appears more basic than the original rock. In the final stages of alteration the last traces of darker color disappear and the entire rock mass becomes white and dense, like clay, and, in many places, shot with sulphur. When wet, such material is plastic. The change in color from dark to light suggests partial reduction and removal of iron, release of silica, removal of much or all of the potassium, sodium, magnesium, and calcium as soluble salts, and conversion of the remaining constituents of the feldspars in part to claylike hydrated aluminum silicates, possibly with magnesium retained. The chemistry of these changes has been discussed by Banfield (1954, p. 771–772).

Origin of the Sulphur

The sulphur is believed to have been precipitated from rising gases by oxidation near the surface (Banfield, 1954, p. 773–774). A similar explanation has been tersely given by Heim (1952, p. 116) for the sulphur in some of the Persian volcanoes. Briefly, part of the hydrogen sulphide gas is first converted by combination with atmospheric oxygen to sulphur oxides (probably SO_2) and water vapor. The sulphur oxides and hydrogen sulphide then probably

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PLATE 3.—AERIAL PHOTOGRAPHS

FIGURE 1.—VOLCAN DARWIN CRATER

Assembled from two photographs, taken from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. *Note* subsidence cliffs and shelf along southwestern side, and lines of craterlets around main crater.

FIGURE 2.—VOLCAN WOLF CRATER

Taken from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. *Note* fringing flows on northeastern and eastern side; also fissure, in part lava filled, on southwest.

PLATE 4.—AERIAL PHOTOGRAPHS OF VOLCAN WOLF

FIGURE 1.—PART OF NORTHERN AND NORTHEASTERN SLOPE

Taken from altitude of about 20,000 feet; approximate scale: 1 inch: 6750 feet. Shows craterlets elongated toward north and breached chiefly in that direction; fissures radiating northward are suggested by alinement of craterlets north-northeast.

FIGURE 2.—NORTHWESTERLY SLOPE

Altitude and scale as in Pl. 4, fig. 1; similar to that view but elongation and alinement of craterlets more westerly; *note* two recent lava flows.

react to form native sulphur and more steam. The process is similar to the Claus method for the production of elemental sulphur from "sour" natural gas. If the natural processes are similar to the artificial one, sulphur from the upward-streaming gases would be deposited only to the depth to which atmospheric oxygen would be available for the chemical reaction. The reaction is exothermic and so may well contribute much of the localized heat.

Significantly, with the exception of the Volcan Chico area, the sulphur-bearing solfataras are not associated with the several lava flows of the present century, such as those on the eastern and southern flanks of Cerro Azul and on the southeastern flank of Volcan Wolf. This suggests that the optimum temperature for sulphur deposition is that of steam, but that in regions where the melting point of these lavas is maintained the sulphur is completely volatilized. Thus a kind of "zoning" results, lava formation, bleaching, sulphur deposition, and steam venting apparently representing successively cooler stages.

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Short Notes

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LATE KEWEENAWAN OR EARLY CAMBRIAN GLACIATION IN UPPER MICHIGAN?

BY WM. C. GUSSOW

Some concern is felt that the contribution by Raymond C. Murray (1955) will be accepted as evidence of pre-Cambrian glaciation. This well-documented and convincing paper describes a most interesting occurrence, and all the evidence supports the conclusion that "moving ice is the most plausible explanation for the features described." The assumption, however, that "this indicates glaciation in upper Michigan in either the late Precambrian or early Cambrian" should not be allowed to stand unchallenged.

Striations, facetted and striated boulders, and polished surfaces are common features along the shores of large bodies of water that freeze in winter. They are not evidence of glaciation and can result from the rafting of seasonal ice by wind or tide. This action grinds off the boulders and pebbles in the frozen beaches and striates the bedrock. These are common features along the shore of Hudson Bay and any northern lake.

If the evidence is truly of glacial origin, it should be a common feature wherever the basal contact of the Jacobsville sandstone is exposed, and there should be regional agreement in strike. The fact that the striations range through 40° supports the conclusion that they were caused by rafting ice. The absence of a glacial till militates against glaciation, and the Jacobsville sandstone itself suggests a near-shore environment for the features described.

Further study of the pre-Jacobsville unconformity may warrant the original assumption that it is in fact evidence of glaciation. But to justify this *a posteriori* conclusion—is the present evidence sufficient?

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LATE KEWEENAWAN OR EARLY CAMBRIAN ICE SHOVE IN UPPER MICHIGAN—A REPLY

BY RAYMOND C. MURRAY

Lake ice set in motion by thermal expansion, wind, or tide is certainly competent under some conditions to produce some of the features cited by this writer (Murray, 1955) as evidence for glaciation. The important question, however, is not the competence of the mechanism but whether or not lake or sea ice produced the features found near L'Anse, Michigan. Let us briefly review and examine the evidence from this exposure with reference to the possibility of striating, polishing, chattermarking, and grooving by rafted lake ice the upper surface of the Michigamme slate where it is overlain by the Jacobsville sandstone.

(1) There is no evidence of lacustrine sediments in the basal arkosic sandstones and conglomerates near L'Anse, and therefore no proof that a large standing body of water existed in the area at this time. Thus the statement by Gussow (1955) that " . . . Jacobsville sandstone itself suggests a near-shore environment" is both misleading and incorrect.

(2) There is no evidence of soled or striated pebbles in the conglomerate beds of the basal Jacobsville in this area that might have acted as tools.

(3) Production of chatter marks requires considerable weight on the cutting tool. Although sufficient thickness of lake ice is presumably possible, it would require extremely unique circumstances. The writer knows of no report of such features from the shores of modern lakes.

(4) There are no extensive exposures of the Michigamme-Jacobsville unconformity. The L'Anse locality is unique in that the upper surface of the slate is locally exposed by removal of the sandstone. Even more remarkable is the production and preservation of the striated surface in this area. If the striations were a local feature produced by lake ice, the coincidence would defy the imagination. The demands made on chance are sufficiently large when this remarkable exposure is considered the product of a more extensive agent such as glaciation.

(5) A 40-degree divergence in strike of the striations is quite commonly produced by glaciers particularly in their terminal areas. This writer recognized two prominent directions at N. 50–60E. and N. 80–90E. Depending on such unknowns as changing wind directions and changes in the size and shape of a given lake, ice could conceivably produce striations with an infinite number of orientations. This argument is inconclusive and suggests only that the features were not produced by a mechanism with a single direction of motion.

In the absence of positive evidence for such minor striation producing mechanisms as mudflows, Pelean clouds, landslides, low-angle meteorites, or ice shove, the writer chose the simplest and most reasonable conclusion in the light of our present knowledge of geological agents. Glaciation, in some form, remains the best explanation for the features.

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